THE SNELLIUS EXPEDITION

IN THE EASTERN PART OF THE EAST INDIAN ARCHIPELAGO 1929-1930

UNDER LEADERSHIP OF
P. M. VAN RIEL †

*

VOLUME II

OCEANOGRAPHIC RESULTS

PART 8

CHEMICAL RESULTS AND A SURVEY OF WATER MASSES AND CURRENTS

BY

H. POSTMA

1958

TO BE OBTAINED FROM THE PRINTERS AND PUBLISHERS
E. J. BRILL — LEIDEN

H. Postma

SNELLIUS-EXPEDITIE

WETENSCHAPPELIJKE UITKOMSTEN DER SNELLIUS-EXPEDITIE

ONDER LEIDING VAN

P. M. VAN RIEL †

LAATSTELIJK DIRECTEUR VAN DE FILIAALINRICHTING TE AMSTERDAM
VAN HET KONINKLIJK NEDERLANDS METEOROLOGISCH INSTITUUT

VERZAMELD IN HET OOSTELIJKE GEDEELTE VAN DE OOST-INDISCHE ARCHIPEL
AAN BOORD VAN H.M. WILLEBRORD SNELLIUS

ONDER COMMANDO VAN

F. PINKE

LUITENANT TER ZEE DER 1e KLASSE

1929–1930

UITGEGEVEN DOOR DE NEDERLANDSE MAATSCHAPPIJ VOOR NATUURWETENSCHAPPELIJK ONDERZOEK VAN OOST- EN WEST-INDIË EN HET KONINKLIJK NEDERLANDSCH AARDRIJKSKUNDIG GENOOTSCHAP

GEDRUKT DOOR EN TE VERKRIJGEN BIJ
E. J. BRILL — LEIDEN

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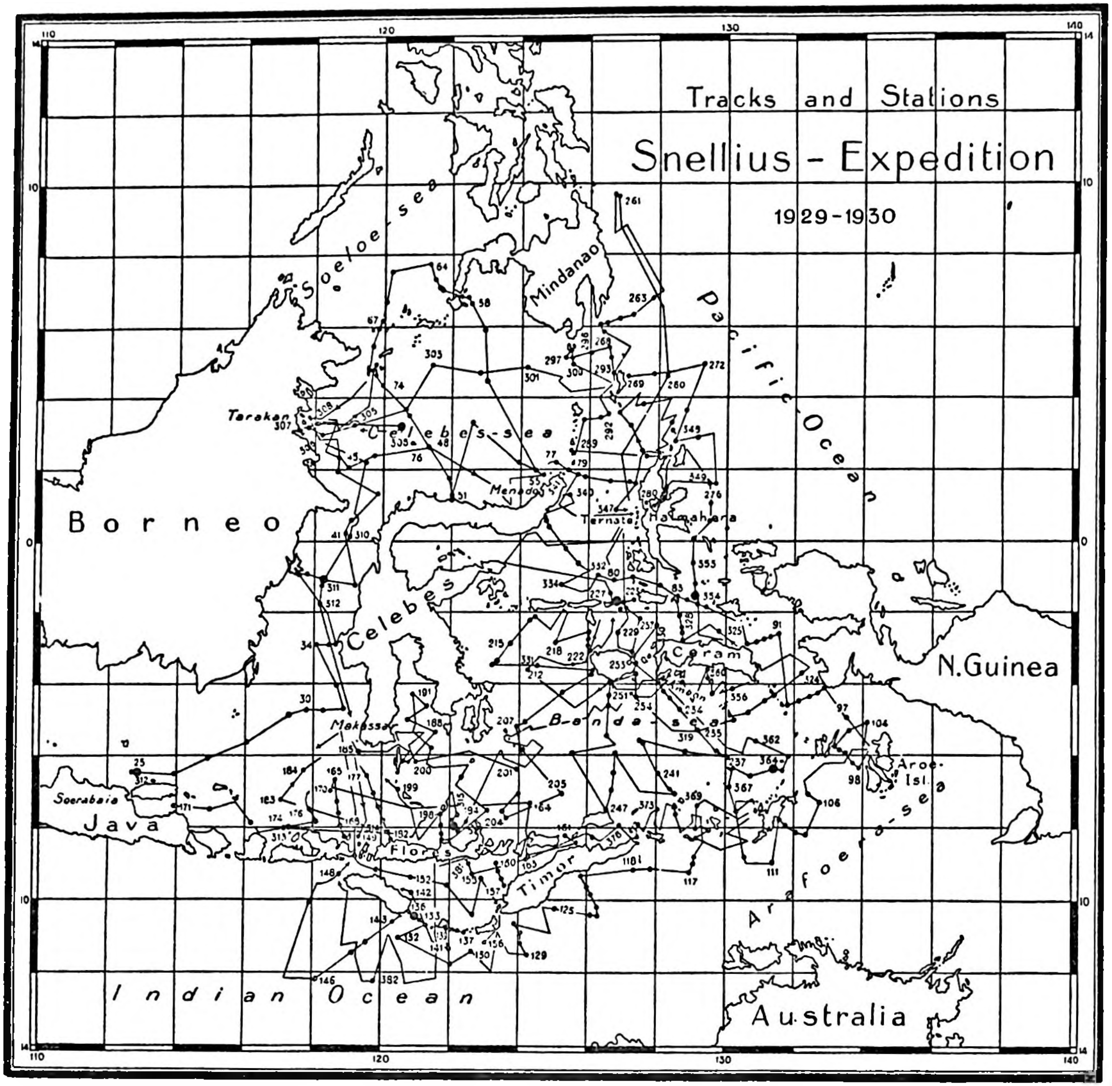

Fig. 1. Tracks and stations of the "Snellius"-expedition, 1929-1930.

CHAPTER ONE

INTRODUCTION

The chemical observations of the "Snellius" expedition comprise measurements of the oxygen content, the hydrogen ion concentration, the alkalinity and the phosphate content of the waters of the East Indonesian Archipelago. Part of the results have been used in previous reports of the expedition for special purposes, especially in connection with the flow of bottom water (VAN RIEL, 1934, 1943; KUENEN, 1943). In the present paper, however, all observations will be discussed as a whole.

In the important contribution to Indonesian oceanography "De Zeeën van Nederlands Oost-Indië" (1922) by VAN DER STOK and others, and in the earlier publications of the "Snellius" expedition the purposes of the chemical measurements were outlined more or less extensively. The observations should form a connection between the hydrographical measurements on the one hand and the biological and geological observations on the other.

The oxygen determinations, together with temperature and salinity, should contribute to a better understanding of the physical oceanography of the area (RINGER, 1922; VAN RIEL, 1937). Both oxygen and phosphate determinations should give insight into the conditions for planktonic and deep water life in the Moluccan seas (BOSCHMA, 1935, VAN RIEL, 1937).

The hydrogen ion and alkalinity data should throw light on the problem of lime solution in the deep basins (TYDEMAN and MOLENGRAAFF, 1922, KUENEN, 1943). Unlike that of the open ocean, the lime content of the basins, first determined by the "Challenger" and "Siboga" expeditions (BÖGGILD, 1916), appeared to be surprisingly low. As tentative explanations for this fact were considered the possibly great capacity for solving lime of the water in the basins, the good ventilation of the basins by convective currents promoted by the heat of the earth's crust and, finally, a relatively abundant supply of non-calcareous material from the surrounding islands.

It follows from the close connection between the chemical and the other investigations, that for a good understanding of the chemical data it will be necessary first to discuss the results of the investigations in the other fields of research. Special attention must be given in this respect to the structure and origin of the water masses occupying the area, the supply of water to the basins, etc. Part of this description can be based on the investigations of other authors, especially VAN RIEL. These investigations must be supplemented, however, by a discussion of data not yet dealt with by previous authors. This holds especially for the distribution of temperature and salinity, and water movement, in the water layers above threshold depths. After a discussion of methods (Chapter II) a separate chapter has therefore been reserved for a discussion of the hydrography of the East-Indonesian seas (Chapter III). Chapter IV, finally, is concerned with the chemical data themselves.

It should be mentioned, that for the chemical measurements carried out the procedures of the German "Meteor" expedition served as examples. The research ship "Meteor" had just returned from her cruises in the Atlantic Ocean when preparations were being made for the "Snellius" expedition. Another point of agreement between the two expeditions is the fact that in both cases a selected part of the oceans was investigated systematically by means of a dense net of oceanographic stations. This method of regional research has the great advantage, that not only a detailed insight is obtained into the oceanography of the part of the ocean in question, but also that on the basis of these regional results conclusions of general value can be drawn.

Two chemists were in charge of the chemical determinations on board of the "Snellius": Dr. A. B. BOELMAN and Dr. H. J. HARDON. The former carried out the measurements of pH, phosphate and alkalinity; the latter was charged with the salinity and oxygen determinations. Mrs. M. M. H. VAN RIEL-VERLOOP took an important part in the oxygen determinations, while members of the

ship's crew carried out the salinity determinations as a routine procedure. All determinations were performed from the beginning to the end of the expedition, with the exception of phosphate, which was measured only during the second part of the expedition. Nitrate determinations, which had been planned in the provisional programme, were not carried out.

The present writer was charged with the compilation of the chemical results many years after the end of the expedition. It is an unfortunate circumstance that the participants of the expedition themselves could not find an opportunity to gather the harvest of two years of continuous and sometimes monotonous effort. For the right evaluation of the chemical results the experience gained during the expedition would have been of the greatest value.

I feel much indebted to Prof. Dr. P. GROEN, assistent-director of the Royal Netherlands Meteorological Institute, and to Dr. J. VERWEY, director of the Zoological Station, for their comments after reading the manuscript. I also wish to express my sincere thanks to Mr. A. D. G. DRAL and Mrs. C. M. PEETERS-PLANKEN for preparation of many of the graphs and charts, and to Mr. J. A. A. SPIEKERMAN for checking the translation into English.

Den Helder, Zoological Station.

CHAPTER TWO

METHODS

The chemical data are compiled in Volume IV of the Reports of the Snellius Expedition (POSTMA, 1958). Table 1 of this volume contains all data, with the exception of phosphate, which is given in table 2. Besides these, table 3 presents data for specific alkalinity, which are obtained by dividing the alkalinity data by the corresponding chlorinities (see II, C). The above-mentioned tables, together with the tables of temperature, salinity and density, published by VAN RIEL, HAMAKER and VAN EYCK (1950) in Vol. II, part 6, comprise all physical and chemical oceanographic results of the "Snellius" expedition, as far as properties of the water are concerned.

In all, the "Snellius" occupied 381 oceanographic stations, 357 of which are located in the Indonesian Archipelago. At nearly all stations samples were taken from the surface to the bottom at small vertical intervals, so that a vast number of observations was collected. Oxygen and hydrogen ion concentrations were determined for nearly all samples. At a large number of stations, however, two series of samples were taken at depths between 0 and 250 m; in these cases mostly only one of the two series was used for chemical analyses. Alkalinity determinations were performed at 123 stations; moreover, the vertical distances between the samples for alkalinity are greater than in the cases of salinity, oxygen and pH, so that a considerably smaller number of observations is available for this property. Phosphate determinations were made at 27 stations during the second part of the cruise, beginning with station 235. Hence only a comparatively small number of phosphate data were collected. Moreover, the stations where phosphate analyses were made are not scattered over the whole area of investigation, but they are restricted to the NE and E parts.

A. *Oxygen.* The method followed for the determination of oxygen and the accuracy of the results have been described by HARDON (1941). All determinations were carried out in duplicate by the WINKLER-method. Although at the first 75 stations the accuracy was somewhat less than afterwards, the reliability of the results was on the whole quite satisfactory.

An exception must be made, however, for the samples taken near the sea bottom with a special bottom water sampler (HAMAKER, 1941, VAN RIEL, 1943). During the first part of the cruise oxygen contents near the bottom were observed which were considerably lower than in the samples taken with the lowest Nansen water sampler, which was suspended on the same wire above the bottom water sampler. At first a decrease of the oxygen content near the bottom seemed quite possible, since consumption of oxygen by organic matter from the bottom could be expected. However, after some time doubt arose whether the metal parts of the bottom water sampler did not consume oxygen themselves. By using a Nansen water sampler at the same depth as the bottom water sampler, it was established that this was indeed the case.

For this reason it seemed advisable to reject the results of the bottom water determinations at the first 200 stations, where the bottom water sampler was always used, the more so as after replacement of this sampler by an ordinary Nansen water sampler a comparable decrease of oxygen near the bottom hardly ever occurred again. VAN RIEL (1943), after a careful discussion of all measurements, nevertheless decided to use the data after a correction of + 0.1 ml/l for all samples obtained with a bottom water sampler from depths exceeding 1500 m.

His arguments for applying this correction were the following. First, the differences in oxygen content between two bottom water samples taken simultaneously at the same depth with a bottom and a normal water sampler systematically increased with depth. The differences were negligible for bottom samples from depths smaller than 1500 m. Below this depth the average difference was 0.1 ml/l. Secondly, after the correction had been applied the stations where still a rapid decrease of

oxygen content near the bottom took place were not scattered over the whole area, but they were located in sharply defined localities in the Sulu Sea, Celebes Sea, Sawu Sea, Timor Trough and Indian Ocean; therefore, this decrease may be a real characteristic of these areas.

At present we can add a third argument to those mentioned above: the observations of the Swedish "Albatross" expedition (1947-1948) have recently proved the existence of a bottom layer with deviating chemical properties over large parts of the ocean floor (Koczy, 1951). Therefore, although from the "Snellius" data no absolute certainty can be obtained about the reality of the phenomenon in the Indonesian basins, the local existence of a bottom layer poor in oxygen seems fairly certain.

A question, also discussed by Van Riel (1943), concerns the fluctuations of oxygen content in the course of the seasons or from one year to another. For this purpose he compared the oxygen determinations from places where observations were repeated with a time interval from six months to a year. The number of stations for which pairs of observations are available is only small and no definite seasonal changes can be deduced from them. The comparison shows, however, that in the upper layers (0-400 m) considerable changes of oxygen content, in some cases up to 0.5 ml/l, frequently occur. In the deeper layers the differences are mostly small with the exception of those in the Moluccan Sea, where it is probable that deep water from the Pacific Ocean has a different southward extension at different times. These changes will be kept in mind when the horizontal and vertical distribution of oxygen is considered (Chapter IV, A).

B. *Hydrogen ion concentration*. For the determination of pH a colorimetric method was followed, which was also used during the "Meteor" expedition (Wattenberg, 1933). A chemical indicator was added to a sea water sample and the colour developed was compared with the colours of a series of buffer mixtures with different pH values prepared according to Palitsch.

A number of corrections should be made to arrive at the actual pH-value in situ [1]). They consist of a salt correction, a correction for pressure, and corrections for the difference between the temperature in situ of the sea water and the temperature of the sample and buffer solution during the measurement (Buch et al., 1932, Wattenberg, 1933, Buch and Nynås, 1939). Table 1 (Volume IV) gives the pH-values after correction for salt error and temperature, but without correction for pressure. Whether the latter should be applied or not depends on the use to be made of the pH measurements. As long as the data are applied to obtain an insight into water movements and the pH-changes caused by biological processes, no correction should be made. However, if one wishes to use pH for ascertaining the equilibria of the carbon dioxide system in situ and the conditions for the solution of lime, the correction should be applied. In this respect a parallel can be drawn between pH and temperature. Also in the latter case it depends on the purpose of the investigation whether one should study the potential temperature or the temperature in situ, i.e. with pressure correction. In the pH-sections presented in this paper the pH values have not been corrected for pressure.

The temperature correction of the pH-values needs some further explanation. At the time when the measurements were carried out, the investigations of Buch et al. concerning the influence of temperature on the value of pH had not yet been published. It was thought that this influence was negligible and consequently no determinations were made of the temperature of samples and buffer solutions during the measurements. Since all measurements were carried out in tropical areas and the temperature of the chemical laboratory was very high, the difference between the temperature of the sea water in situ and that during the pH-determination will often have been considerable, especially for bottom water samples. After consultation of Mr. Van Riel the average temperature of the ship's laboratory was estimated at 28°C and all corrections have been made on the basis of this temperature. A similar procedure was followed by Wattenberg (1933) for the pH-measurements of the "Meteor" expedition. There remains some uncertainty whether all samples had actually reached the temperature of the laboratory before the pH measurement was carried out. We should add, however, that probably the water samplers themselves, when drawn up to the surface, already attained a temperature close to 28°, since in the upper water layers, down to about 100 m, the temperature was always above 25°C.

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[1]) The pH-values published in earlier reports (Van Riel, 1943, Hamaker, 1941) are uncorrected.

In the case of colorimetric determinations it is difficult to keep measurements of different series comparable with a sufficient degree of accuracy over so long a period as 2 years. A comparison of pairs of stations from the same locality, occupied with an interval of 6 months or more, shows, that the pH-values during the first part of the cruise are indeed mostly somewhat higher than during the latter part (table 1).

Table 1

Difference of hydrogen ion concentration of pairs of stations from the same place, occupied with an interval of 5 months or more. Averages along the vertical. Δ pH negative: decrease of pH.

Stations	Time interval	Δ pH
39-311	1 year	— 0.003
41-310	1 year	— 0.06
75-304	9 months	— 0.04
79-342	1 year	— 0.007
80-332	1 year	— 0.006
84-329	1 year	— 0.065
155-379	1 year	— 0.02
212-331	6 months	+ 0.04
235-319	5 months	— 0.05

The pH-changes in the course of time are not systematical enough to allow a correction for this methodical error. However, they are on the whole very small. When preparing the vertical sections of pH-distribution, it nevertheless seemed advisable to make use only of stations which were occupied within a comparatively short period. For this reason, the stations used for the composition of vertical sections of pH are not always the same as those used for the salinity and temperature distribution.

The absolute accuracy of the pH-values measured can be checked to a certain degree by comparing the "Snellius" measurements with those of the more recent "Albatross" expedition (Bruneau, Jerlov and Koczy, 1953). During this expedition the hydrogen ion concentration was measured by means of a Beckman pH-meter, model G. The reliability of such an apparatus for pH-measurements in sea water was already established by Buch and Nynås in 1939. One of the "Albatross" stations (nr. 162) in the Pacific Ocean occupies a position sufficiently near to the "Snellius" station nr. 270 to allow a comparison between the two series of pH-measurements (table 2). At depths between 150 and 800 m the "Snellius" measurements are higher than those of the "Albatross", whereas at all other depths the reverse is the case. The differences only once exceed 0.05 pH-units. The average pH-difference between the two series is practically nil. The comparison is, therefore, very satisfactory, the more so if one takes into account that the two expeditions are separated by a period of about 25 years.

Special attention must finally be given to the pH-measurements of the bottom water. Hamaker (1941) already concluded that the bottom water sampler used during the first half of the expedition not only consumed oxygen, but also caused a decrease of the hydrogen ion concentration. For the stations where the bottom sampler was used, the average pH-difference between the bottom sample and the sample taken with the lowest ordinary water sampler amounted to 0.15 pH-units. In a number of cases even a difference of about 0.50 pH-units was measured. At four stations the pH was measured at the same depth simultaneously with an ordinary sampler and a bottom water sampler (table 3). The average pH-difference of these samples amounted to 0.15 pH-units. For the stations where the bottom sampler was not used, the increase of pH near the bottom was practically nil. In table 1 (Volume IV) the pH-observations in question have been corrected by subtraction of 0.15 pH-units.

Table 2

Comparison of hydrogen ion concentrations measured by the Swedish "Albatross" expedition (1947-1948) and the "Snellius" expedition. Data uncorrected for pressure.

Depth	Albatross, St.162 5°23′ N, 27°48′ E	Snellius, St. 270	Δ pH
0	8.16	8.17	+ 0.01
50	8.17	8.15	— 0.02
75	8.12	8.12	0.00
100	8.09	8.05	— 0.04
150	7.97	8.00	+ 0.03
200	7.91	7.99	+ 0.08
250	7.91	7.96	+ 0.05
300	7.88	7.93	+ 0.05
400	7.83	7.87	+ 0.04
500	7.81	7.84	+ 0.03
600	7.80	7.81	+ 0.01
790	7.77	7.79	+ 0.02
990	7.81	7.76	— 0.05
1450	7.81	7.76	— 0.05
1800	7.81	7.78	— 0.03
2400	7.85	7.83	— 0.02
3000	7.89	7.89	0.00
4000	7.90	7.90	0.00
4700	7.92	7.89	— 0.03

After this correction there remain a number of stations where a significant increase of pH near the bottom in comparison with the lowest serial observation still occurs. Table 4 lists all stations where this increase is greater than 0.10 pH-units. Just as in the case of oxygen these stations lie in closely defined localities of the Sulu Sea, Celebes Sea, Makassar Strait and Arafura Sea. The possibility cannot be excluded, therefore, that the pH-increase is real, the more so as a pH-increase close to the bottom was also found by the "Albatross" expedition (Koczy, 1951) in the open sea.

Table 3

Comparison of pH of samples from the same depth obtained simultaneously with different water bottles.

Station	Bottom water sampler	Ordinary serial sampler	Δ	Depth, m
197	8.02	7.80	0.22	5100
200	7.90	7.80	0.10	1065
205	7.95	7.80	0.15	3903
264	7.91	7.78	0.13	± 4200
		Average	0.15	

Table 4

Stations with an increase of pH in the bottom water greater than 0.10 pH-units after correction for the influence of the bottom water sampler.

Station nr.	Bottom depth, m	Bottom sample depth, m	Increase of pH
33	1932	1901	0.25
42	690	659	0.11
48	5515	5484	0.12
49	3010	2979	0.20
50	1713	1682	0.11
53	5004	4973	0.15
58	2722	2691	0.17
64	4299	4268	0.15
74	2637	2606	0.14
75	4773	4742	0.12
76	5544	5513	0.15
80	4617	4568	0.23
100	3594	3563	0.11
106	714	683	0.11

C. Alkalinity. The determination of "excess base" was carried out according to the titration method of WATTENBERG (1933). For this determination 10 ml 0.1 n hydrochloric acid were added to a sea water sample of 200 ml and the excess amount of acid was titrated back with 0.05 N barium hydroxyde solution; the indicator used was a mixture of bromcresolgreen and methylred. The results of the determinations, expressed in milli-equivalents per litre, are compiled in table 1 (Vol. IV) Moreover, table 3 (Vol. IV) gives the same data in the form of specific alkalinities. The specific alkalinity of a sample is calculated from the alkalinity by dividing the latter by the corresponding chlorinity. As a result specific alkalinity differences between samples are independent of salinity differences. In fact it would have been more exact to divide alkalinity by chlorosity (grams of chlorine per litre) instead of chlorinity (grams per kilogram), thus using units with the same dimensions. However, the difference is mainly formal and by the use of chlorosity a direct comparison with WATTENBERG's results would have been difficult.

For the determination of the reliability of the estimations we cannot, as in the cases of oxygen content and pH, compare the results of stations occupied in the same place but at different times during the expedition, since the alkalinity was never determined twice in the same locality. Nevertheless, there is much evidence that the alkalinity determinations of the first part of the expedition are too high in comparison with those of the latter part. However, to elucidate this fact we have to discuss in detail the alkalinity distributuon and this discussion will better be postponed until after the complete review of alkalinity data in chapter IV, B.

In this respect an exception must be made for the alkalinity measurements in the bottom water. Exactly as in the cases of the oxygen and pH measurements we have to consider the possibility that the bottom water samplers used during the first part of the expedition influenced the alkalinity values. Unfortunately, no direct means of comparison is available, since no duplicate samples were taken with a bottom water sampler and a normal Nansen water sampler. Inspection of table 3 (Vol. IV) shows, however, that at eight stations (76, 79, 80, 95, 99, 144, 147 and 149) abnormally high specific alkalinities of the bottom water were found. Two of these stations are located in the Celebes Sea, one in the Ceram Sea, one in the Flores Sea, two in the Arafura Sea, and two in the Indian Ocean. The distribution of these stations is therefore quite different from those with deviating oxygen and pH values. This is one reason to consider these observations with some doubt. A second reason is,

that in the Arafura Sea and the Indian Ocean the bottom water at the depths of the observations is oversaturated with lime (see Chapter IV). This seems to exclude the possibility of a substantial solution of calcium carbonate from the ocean bottom. In the third place, during the latter part of the cruise a considerable increase of specific alkalinity never occurred again. Taking these facts together, it seems advisable to consider these values as erroneous ones.

D. Phosphate (table 2, Volume IV). For the measurement of this nutrient the well-known method of DENIGÈS was used, which was introduced into marine chemistry only a few years before the beginning of the "Snellius"-expedition by ATKINS. To correct for salt error the original values have been multiplied by 1.3 (WATTENBERG, 1937).

The accuracy of the absolute phosphate-values found can be checked to some extent by comparison with those of the "Dana"-expedition, which occupied a considerable number of stations in the Indonesian seas in 1929-1930. Besides, one station of the "Albatross"-expedition (nr. 162) is located sufficiently near to the "Snellius"-stations in the Pacific Ocean to allow a comparison.

The "Dana" phosphate determinations in the deep water are on the average 0.4-0.5 μg-at/l higher than those of the "Snellius"-expedition, as can be seen from figs. 14 and 15 (p. 61).

The measurements at the "Albatross"-station nr. 162 (table 5) are 0.5-0.6 μg-at/l higher than the "Snellius"-observations in the vicinity. Hence there can be little doubt that the "Snellius"-determinations are much too low. The relative accuracy, however, may be sufficient to allow a comparison of the estimations in different areas among themselves.

Table 5

Comparison of phosphate measurements of the "Albatross" (1947-1948) and "Snellius" expeditions. Data corrected for salt error.

Depth	Albatross, St. 162 PO_4, μg-at/l	Snellius, St. 272 PO_4, μg-at/l	Δ PO_4
0	0.19	0.35	— 0.16
100	0.93	0.45	+ 0.48
340	2.12	1.61	+ 0.51
900	2.81	1.92	+ 0.89
1900	2.96	2.60	+ 0.36
2900	2.78	2.21	+ 0.57
3900	2.71	2.29	+ 0.42
4900	2.78	2.18	+ 0.60

To conclude the discussion of this chapter, table 6 summarizes the results of the reliable multiple observations near the bottom; all stations with a doubtful accuracy have been omitted (nrs. 66, 76, 80, 104, 145, 163, and 180). The table once more stresses the fact that, as soon as the observations of the first part of the cruise are left out of account, no important changes in the concentrations of the various substances close to the bottom appear to occur.

Table 6

Multiple observations close to the bottom.

Station, nr	Depth, m	Hydrogen ion, pH [1])	Specific Alkalinity, A/Cl	Phosphate, μg-at/l [2])	Oxygen, ml/l
198 (2804m)	2750 2771	7.81 7.82	0.1279 0.1296	— —	2.28 2.26

Station, nr	Depth, m	Hydrogen ion, pH [1]	Specific Alkalinity, A/Cl	Phosphate, μg-at/l [2]	Oxygen, ml/l
202 (3898m)	3805	7.77	—	—	—
	3835	7.81	0.1288	—	2.39
	3865	7.79	—	—	—
205 (3936m)	3843	7.96	—	—	—
	3873	7.81	0.1290	—	2.45
	3903	7.80	0.1290	—	2.36
209 (4224m)	4142	7.84	0.1291	—	2.53
	4172	7.85	0.1296	—	2.43
	4202	7.83	0.1292	—	2.53
212 (4968m)	4898	7.81	—	—	2.46
	4928	7.81	—	—	2.45
	4958	7.82	—	—	2.46
253 (4024m)	3931	7.78	—	—	2.31
	3961	7.79	—	—	2.44
	3991	7.80	—	—	2.42
278 (474m)	400	7.79	0.1225	1.55	0.00
	431	7.81	0.1241	2.07	0.00
	441	7.77	—	—	0.00
	464	7.78	0.1247	2.30	0.00
309 (5108m)	5015	7.77	0.1268	—	2.08
	5045	7.76	0.1265	—	2.05
	5075	7.76	0.1265	—	2.08
330 (4450m)	4360	7.83	0.1289	2.69	2.55
	4390	7.82	0.1281	2.75	2.54
	4420	7.81	0.1290	2.38	2.55
331 (5026m)	4476	7.87	—	—	2.48
	4726	7.84	—	—	2.50
	4976	7.83	—	—	2.50
362 (7326m)	7233	7.81	0.1288	2.18	2.36
	7263	7.82	0.1280	2.52	2.35
	7293	7.81	0.1288	2.22	2.38

[1]) uncorrected for pressure
[2]) corrected for salt error

CHAPTER THREE

WATER MOVEMENT AND DISTRIBUTIONS OF SALINITY AND TEMPERATURE

The position of the Indonesian seas and basins between two oceans (figs. 2 and 3) raises the problem from what sources the water masses in these areas originate. This question has partly been discussed in earlier papers of the "Snellius-expedition". VAN RIEL, in a series of contributions (1934, 1943, 1956), traced the origin of the deep water in the basins and discussed the oxygen content of the water and the temperature distribution in the deep water. KUENEN (1943), in connection with the sedimentology of the area, speculated about the mechanism and the rate of water renewal in the basins. LEK (1943) discussed the current measurements and VISSER (1938) treated the distribution of salinity and temperature in the surface waters.

The insight into water movement and water renewal obtained from the work of the above-mentioned authors is by no means complete, however, since the three-dimensional distribution of temperature and salinity has hardly been taken into account, especially for the layers between the surface and the thresholds of the basins. The present chapter intends to fill this gap. It is quite obvious that this knowledge will be of primary importance for a good understanding of the chemistry of the area.

By means of the salinity and temperature data, published by VAN RIEL, HAMAKER and VAN EYCK (1950) a number of graphs have been constructed, which represent the main features of the vertical and horizontal distribution of these values (figs. 16 and 17 A-G, pp. 65-74). The traverses followed by the vertical sections are indicated in fig. 2; they follow courses over the deepest parts of the thresholds.

It appeared convenient to divide the discussion into three parts. The first part will be concerned with the distribution of temperature and salinity and the water movements in the layers between the surface and the thresholds; in the second part the conditions of the water masses inside the basins will be discussed. The third part, finally, gives a review of water movement as a whole by means of temperature-salinity relations.

A. *Water movements above the thresholds.* The number of unobstructed passages in the Indonesian Archipelago and between the Pacific and the Indian Ocean decreases with increasing depth, and the thickness of the layer in which a free flow of water from one region to the other is possible depends on the depth of the passage in question. Therefore, the lowest boundary of the water masses to be discussed in this section cannot be defined exactly. However, the most important lower limit is certainly the depth of about 1400 m at which the last remaining free connections between the Pacific and the Indian Ocean are closed. One of these connections is located east of the island of Timor, the other forms the northern entrance to the Aru Basin. The water masses in the levels above 1400 m can therefore in most parts of the area originate from the Pacific as well as from the Indian Ocean.

To obtain some sort of impression of the structure of the water masses, five stations have been selected to show the vertical distribution of temperature, salinity, density and, for the sake of completeness, oxygen (fig. 4). Down to about 200 m, the temperature decreases rapidly with depth and the density shows a similar increase. At a depth of 100-150 m a salinity maximum is present, which is especially well developed in the two Pacific stations 276 and 296; it is somewhat less pronounced at station 31 in the Strait of Makassar and practically absent in the two southern stations 144 and 235. Below the salinity maximum, the stations show a salinity minimum at intermediate depths.

In view of this vertical stratification, the discussion of the conditions above the thresholds will be divided into three parts. First, the surface layer will be considered; next, attention will be given

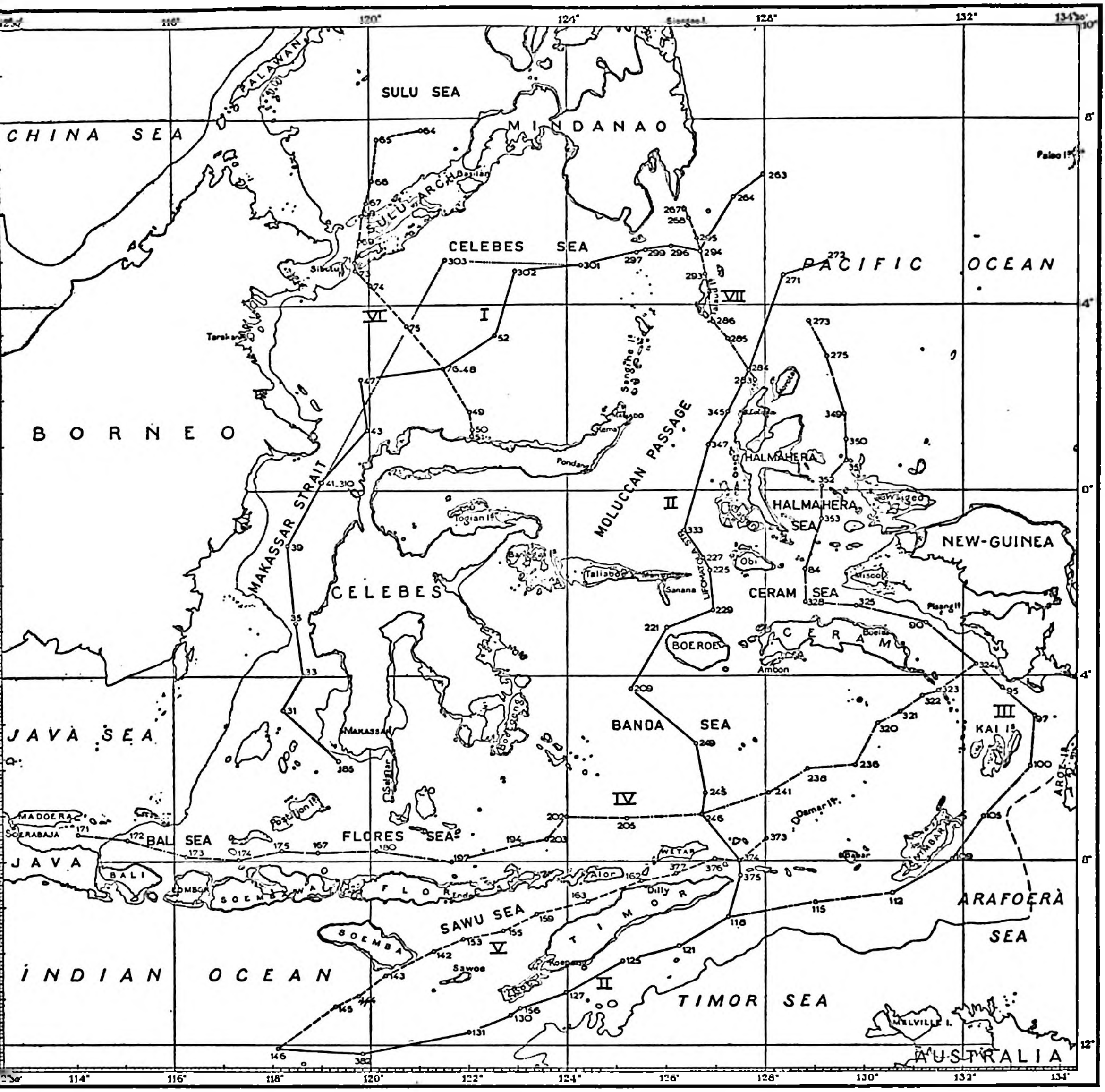


Fig. 2. Principal seas and sections I-VII, used to represent the vertical distribution of temperature and of chemical substances (figs. 16-20); the dotted line indicates the 200 m depth contour.

to the water masses from the surface down to and including the salinity maxima; finally, the conditions in the intermediate water masses will be discussed.

1. In the surface layers the direction of water movement is not constant the whole year round, but it depends very much on the direction of the winds, which change with the seasons. During the dry season, from about June to about November, the winds in the East Indonesian Archipelago generally blow from the SW north of the Equator and the SE south of the Equator. In the months of December-May the directions are NE and NW, respectively. The currents change correspondingly,

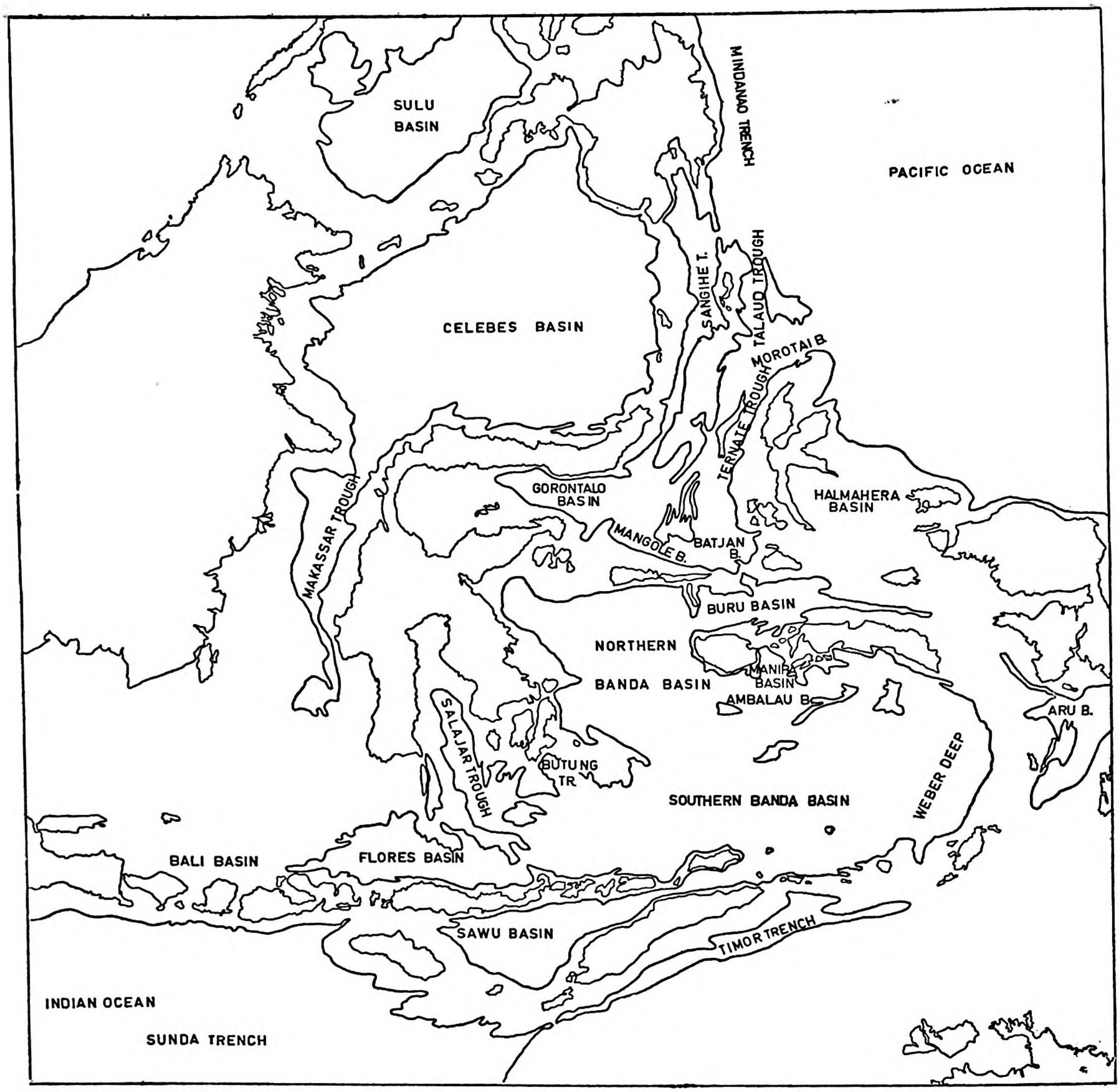


Fig. 3. Principal basins, indicated by the 2000 m depth contour line.

so that for example from December to May water flows from the shallow Java Sea into the area studied, whereas in the other half of the year the reverse occurs. Through the shallow Torres Strait between New Guinea and Australia Pacific water enters from June to November. Details of winds and currents can be obtained from the meteorological and oceanographical atlasses of the Royal Netherlands Meteorological Institute (1935, 1936, 1949), which give monthly averages of velocities and directions in one or two degree squares (see also Van der Stok, 1922 and Lek, 1943).

In certain parts of the area, however, a current in one direction is maintained the whole year round, although the strength of the current varies with the seasons. This holds in the first place for the current flowing through the areas east of Mindanao, the Celebes Sea, the Makassar Strait and the

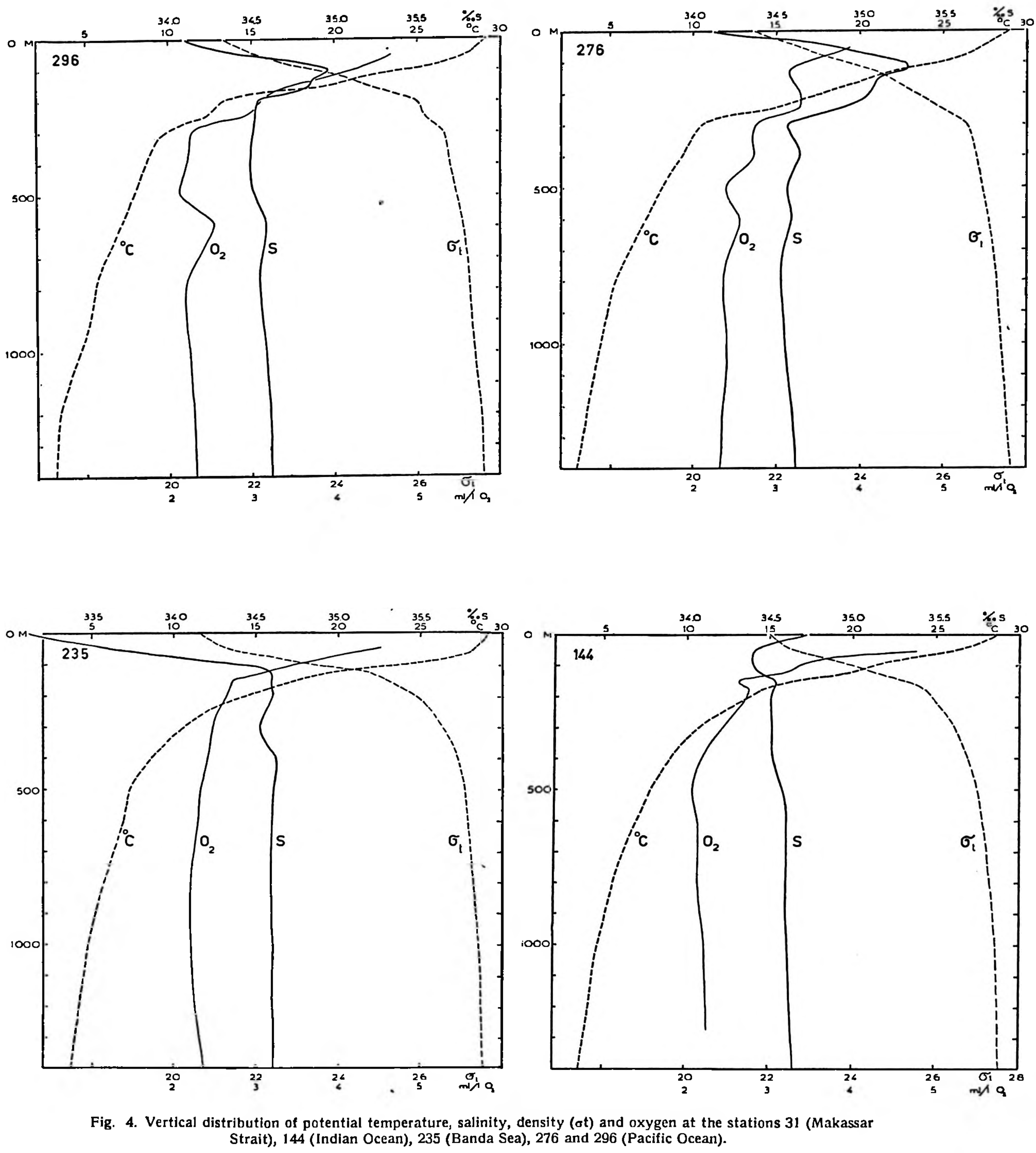


Fig. 4. Vertical distribution of potential temperature, salinity, density (σt) and oxygen at the stations 31 (Makassar Strait), 144 (Indian Ocean), 235 (Banda Sea), 276 and 296 (Pacific Ocean).

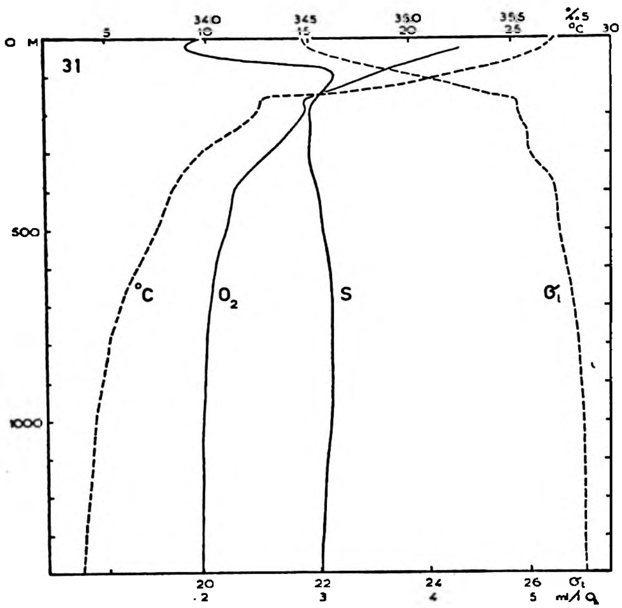


Fig. 4 (continued)

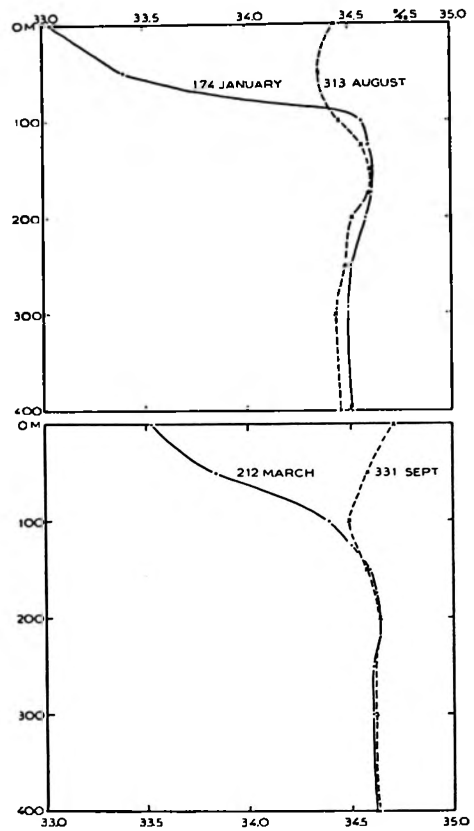


Fig. 5. Seasonal variations of salinity in the surface layers at the station pairs 174 and 313 (Flores Sea), and 212 and 331 (N Banda Sea).

Flores Sea. This current is a continuation of the Pacific North Equatorial Current, which flows along Mindanao to the south (Mindanao Current). Also in the Halmahera Sea the current is in the greater part of the year directed away from the Pacific Ocean. From June to November this current forms a branch of the South Equatorial Current, which, north of New Guinea is well developed during this time of the year. In the other half of the year the whole region between Mindanao and New Guinea is occupied by water from the Mindanao Current (SCHOTT, 1939).

In the Molucca Sea the surface currents flow alternately from and towards the Pacific Ocean with a predominance of the latter direction. This area may therefore more of less be considered as the root of the Equatorial Counter Current.

The influence of the Pacific Equatorial currents has the result, that on the whole more surface water flows from the Pacific Ocean into the Indonesian seas than in the reverse direction. The reason for this Pacific predominance may be explained as follows. It has been shown among others by MONTGOMERY and PALMÉN (1940) that surface water is piled up by the trade winds against the western side of the Pacific Ocean. Most of this water returns to the east in higher latitudes or with the Equatorial Counter current. The piling up of water, however, may also cause a slope from the Pacific Ocean to the Indian Ocean. From this point of view the Indonesian passages form, so to speak, a leak in the western boundary of the Pacific Ocean, through which a certain amount of water escapes towards the Indian Ocean.

The salinity of the surface layers changes in the course of the year, due to changes in the direction of the currents and variations in precipitation and river run-off. These variations have been studied by VISSER (1938) and, more recently, by VEEN (1952) and WYRTKI (1956).

The wet monsoon (December-May) carries water of low salinity from the Java Sea and the Makassar Strait to the east. The rivers of New Guinea cause a decrease of salinity around this island. The salinity only locally drops below 30‰, however. Values above 34‰ occur in the Banda Sea and in the Pacific and the Indian Ocean. In the dry season (June-November) the surface salinity is practically in the whole region above 34‰.

A comparison of wet and dry season data from the same area shows, that the monsoon influence on the salinity of the surface layers penetrates as far down as about 100 m. (fig. 5).

2. The movements of the water masses in the layers immediately below the surface can be very clearly demonstrated by means of the salinity distribution. The vertical sections of salinity (fig. 16, sections I, II, and III) show that at a depth of 100-200 m water with a high salinity enters the Indonesian seas from the direction of the Pacific Ocean. During the transport towards the west and south the salinity maximum becomes gradually less pronounced and practically disappears in the southern and central areas, including the adjacent part of the Indian Ocean.

In the Pacific Ocean itself a salinity maximum is present over very large regions; in fact two sources of water of high salinity can be distinguished, one located 20-30° north and the other about the same distance south of the Equator (Schott, 1935, Sverdrup et al, 1946). The saline water formed in these areas at the surface sinks down to about the bottom of the upper mixed layer. The northern water, carried to the west by the North Equatorial Current, has a lower salinity and slightly higher temperature than the southern water, which is transported by the South Equatorial Current. Both water masses can still be distinguished by their difference in salinity in the area between Mindanao and New Guinea. The highest salinities, with values above 35.20‰, occur in the vicinity of the latter island (section III).

The question arises where the boundary between the two water masses is to be found. It should at once be observed that this boundary will shift with the seasons, since in the period from November to May the South Equatorial Current becomes insignificant in this part of the Pacific. The "Snellius"-observations were made here in the other half of the year, when the current is well developed. Inspection of the salinity data reveals the presence of a double salinity maximum at stations north of the island of Halmahera. A number of these stations are reproduced in fig. 6, together with the

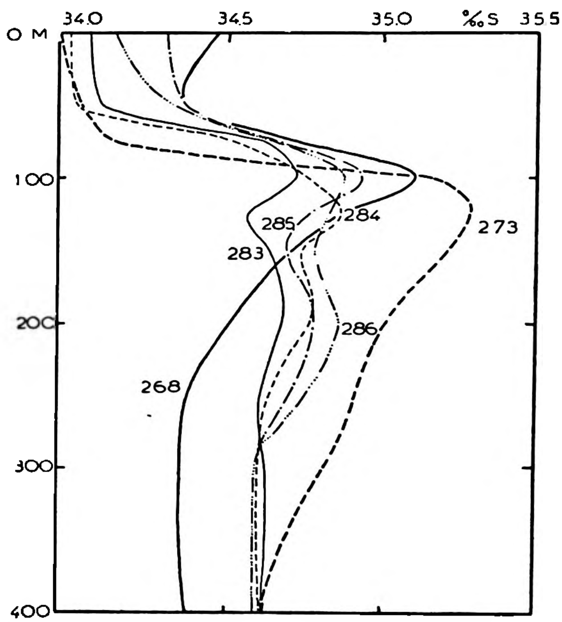


Fig. 6. Stations in the northern Molucca Sea with a double salinity maximum; the Pacific North Equatorial salinity maximum is represented by station 268, the South Equatorial maximum by station 273.

stations 268 and 273, representing the conditions in the North and South Equatorial Current respectively. Evidently during the half of the year under consideration the boundary between the two currents lies north of Halmahera and east of the Talaud islands; according to fig. 6 in the transition area the northern salinity maximum probably lies above the southern one.

A second means to establish the position of the boundary is the distribution of oxygen, to be discussed in Chapter IV. In anticipation of this discussion it may already be mentioned here, that the oxygen content of the North Equatorial Current is higher than that of the South Equatorial Current. In this manner it can be proved that the Molucca Sea is filled with northern water, while the above-mentioned area again has an intermediate character.

Although the Molucca Sea, according to the facts mentioned above, is evidently filled with northern water, it nevertheless occupies a particular position. Whereas the layer of maximum salinity can be followed from the Pacific Ocean very far into the Celebes Sea and the Halmahera Sea, it rapidly becomes indistinct already in the northern Molucca Sea. This may indicate that the Molucca Sea has more or less the same character as the Equatorial Counter Current, where the salinity maximum is also absent, certainly because of an ascending water movement. In this sense the Molucca Sea water, just as the Equatorial Counter Current, may occupy an intermediate position between the North and South Equatorial Current. The Equatorial Counter Current itself, according to SCHOTT (1939), has its origin east of the Talaud islands.

3. Below the layer of maximum salinity, the salinity in the Pacific Ocean gradually decreases down to a certain depth and then increases again. South of the Talaud islands (sections II, III and S. part of section VII) the salinity minimum lies at a depth of about 800 m. North of these islands (section I and N. part of section VII) a minimum is present at a depth of about 350 m. This minimum has an irregular appearance and is at one station better developed than at others.

The minima gradually become less pronounced in a direction away from the Pacific Ocean; they therefore certainly have their source in this ocean. The minimum at 800 m, which is about 34.55$^0/_{00}$ and is accompanied by a temperature of about 5.5°C, must originate from the Antarctic Intermediate Water. The latter water mass is formed at the surface in sub-antarctic regions, where it has a temperature between 3 and 7°C and a salinity below 34$^0/_{00}$; it gradually descends to a depth of 800-1000 m and at the same time moves in a northern direction (SVERDRUP et al., 1946). During this transport the salinity increases as mixing with water masses from above and below takes place, until near New Guinea the value mentioned above is attained. The temperature remains approximately the same.

The minimum at 350 m is slightly below 34.40$^0/_{00}$ and the water has a temperature of about 10°C. This minimum contains surface water from the North Pacific Ocean, where above 45°N a sub-arctic water mass is formed with a low temperature (2-4°C) and a surface salinity as low as 32$^0/_{00}$, which flows along the American coast to the south and in lower latitudes to the west (SVERDRUP et al., 1946). During this transport the temperature and the salinity in the upper layers increase owing to heating and evaporation and in the deeper layers intrusion occurs of salt water from below. These processes result in the formation of the salinity minimum at 350 m, which near Mindanao has the properties mentioned above.

The area between New Guinea and the Philippines evidently not only forms a meeting-place of the North and South Equatorial Currents, but also of the intermediate water masses of the North and South Pacific Ocean (to be called hereafter North and South Pacific Intermediate Water, respectively). Attention to this fact was already drawn by MEGIA and VILLADOLID (1953). Apparently, the northern water in one period penetrates further to the south than in another.

We will now follow the distribution of the two salinity minima in the Indonesian seas. The South Pacific Intermediate Water enters the area through all passages, but in those north of the Talaud islands, and also in the Celebes Sea, the salinity minimum of this water mass is eclipsed by the more pronounced minimum of the North Pacific Intermediate Water. Only at a few stations both minima can be discerned, for example at the stations 52 and 264 of section I (fig. 16 A), and also at station 53 in the Celebes Sea and at the stations 291 and 292 SW of the Talaud islands. An inspection of all salinity data shows that no stations with a salinity minimum at 350 m are found SE of the line connecting the Talaud islands with the NE point of Celebes.

The northern salinity minimum, which enters the Indonesian area close to the island of Min-

danao (section VII) spreads from this entrance over the whole Celebes Sea from north to south. This is confirmed by the salinity distribution of section VI. Sections I and IV show, that the northern salinity minimum is still very well marked in the Makassar Strait and even in the Flores Sea after passing over the sill between these two areas. The salinity minimum is present not only in the Flores Sea, but also farther to the east in the whole of the Southern Banda Sea. Its influence extends here to greater depths. In the Sawu Sea (section V) the minimum is present at about 400 m, which seems to prove that much of the Flores Sea water passes through this area to the Indian Ocean.

The salinity distribution along section II indicates that the southern salinity minimum at 800 m enters the Molucca Sea and is well marked here until the water passes through the Strait of Lifomatola. Since the vertical differences of salinity are largely levelled out here, intensive mixing processes appear to occur in this narrow passage.

As a result, the salinity minimum has disappeared in the water entering the Ceram Sea and the northern Banda Sea. Another feature of the salinity distribution in these areas attracts our attention, however. As section II shows, water of comparatively high salinity is present at a depth of 250-600 m in the Buru Sea (western Ceram Sea, stations 225 and 229). This salinity maximum must originate from the Halmahera Sea, the only adjacent area where comparable salinity values occur (section III).

Hence, a westward current must dominate water movement in the Ceram Sea between 250-600 m. The Banda Sea, therefore, in these layers evidently receives much water from the Halmahera Sea. Below 600 m the connection between the Ceram Sea and the Halmahera Sea is interrupted and the only remaining connection with the Pacific Ocean is formed by the Strait of Lifomatola. In the deeper layers of the Ceram Sea the direction of flow must therefore be opposite to that in the layers above 600 m.

We conclude from the above discussion, that the Banda Sea receives Pacific water at all depths below the surface layer and through all passages. This fact, which is also confirmed by the current measurements of the "Snellius" in these passages, implies that water must leave the Banda Sea towards the Indian Ocean. Nevertheless the possibility remains that in the southern seas of the Indonesian Archipelago water is also supplied from the Indian Ocean. In fact the current measurements show that through the passage between the Sawu Sea and the Indian Ocean intermediate water probably flows to the north (LEK, 1938, anchor station 135). One might for example suppose, therefore, that Indian Ocean water enters the Sawu Sea, flows into the Banda Sea and returns to the Indian Ocean through the Timor Sea.

The distribution of salinity and temperature does not provide sufficient evidence to prove or disprove this possibility. However, anticipating the discussion in Chapter IV, we may already state here, that the oxygen distribution does exclude the possibility of a substantial supply of Indian Ocean water. It will be shown in this chapter that the water mass above 900 m in the Indian Ocean near the area investigated consists for the greater part of Indonesian water with only a slight admixture of Indian Ocean water. Water masses containing a substantial amount of Indian Ocean water occur only in the Timor Sea below the above-mentioned depth.

In the following pages this conclusion will, where necessary, already be taken into account.

When the discussion proceeds it will be shown that the most important characteristic of the East Indonesian Archipelago is, that it forms the meetingplace of Pacific waters from different origin. *Moreover, it will become increasingly clear that the flow of Pacific water along the route Celebes Sea-Makassar Strait-Flores Sea-Banda Sea and from there towards the Indian Ocean is by far the most important water transport quantitatively.* This assertion is based not only on the salinity data already discussed and the current measurements already mentioned, but it will also be strongly supported by the temperature distribution, the TS-relations, the oxygen distribution and, in some extent, by the distribution of other chemical values.

A description of the distribution of temperature and salinity in *horizontal sections* may add a number of details to the previous discussion which have escaped attention so far. Successively, the distributions at 50, 100, 200, 400, 600, 800, 1000 and 1250 m will shortly be examined. Greater depths will not be considered, since below about 1400 m the conditions are wholly determined by the depths of the entrances to the various basins.

50 m (figs. 21 A and 22 A, pp. 88 and 96).

The seasonal influence on the salinity distribution, which is strongly felt at the surface, is also well marked at 50 m; for this reason the observations for this depth must be divided into those for the dry and those for the rainy season. Only for the former period (June-November) are sufficient data for the construction of a chart available.

In many respects the salinity distribution at 50 m resembles that at the surface (VISSER, 1938). Saline water of the Pacific South Equatorial Current enters the area between New Guinea and Halmahera and can be traced as far as the Arafura Sea and the northern Banda Sea (isohaline of 34.55 ‰). Water with a comparatively low salinity of the North Equatorial Current penetrates into the Celebes Sea and the Molucca Sea. In the Strait of Makassar this water is diluted by less saline surface water from near the coast of Borneo and the Sulu Sea. From there the water flows into the Flores Sea, where part of it enters the Java Sea, whereas another part is transported into the Southern Banda Sea. As a result there is a significant difference in salinity between the northern and southern Banda Sea.

The temperature distribution shows a number of small and isolated places with a comparatively low temperature and one identical larger area in the eastern Banda Sea and the Arafura Sea. In these areas water from deeper layers evidently comes nearer the surface, either as a result of vertical mixing or by processes of upwelling.

100 m (figs. 21 B and 22 B).

Although seasonal variations also exert their influence at 100 m, these are comparatively small, and all observations can therefore be used together in the same chart, without regard to the time of the year.

The salinity values are higher than those at 50 m, especially near Mindanao; this increase is due to the nearness of the salinity maximum, which near Mindanao lies at a depth of 100-150 m and near New Guinea at 140-160 m. As a result the salinity in the Celebes Sea is much higher at 100 than at 50 m. The difference is even greater in the Strait of Makassar, where the fresh water influence, which is strongly felt at 50 m, hardly penetrates into deeper layers. Hence, also the salinity of the Flores Sea and the southern Banda Sea has increased considerably; the horizontal differences in salinity within the Banda Sea are comparatively small over large distances.

The area with a comparatively low temperature in the eastern Banda Sea, observed at 50 m, also appears at 100 m. A new area of low temperature, NE of the Talaud islands, is even more clearly marked at 200 m and will therefore be discussed below.

200 m (figs. 21 C and 22 C).

Also at this depth the highest salinity values occur in the area near New Guinea; the Celebes Sea water not only has a lower salinity, but also a lower temperature. The differences between these two water masses are maintained when they enter the Banda Sea. As a result the Celebes Sea-Makassar Strait water passing through the Flores Sea into the Banda Sea can readily be distinguished from the water entering the Banda Sea from the north. It is remarkable that much of the water entering the Banda Sea via the route Celebes Sea-Makassar Strait-Flores Sea does not follow the shortest course towards the Indian Ocean, but flows at first far towards the East. The sections III and IV (figs. 16 C and D) show this Flores Sea water even to be present in the area near New Guinea. This does not only hold for the water layer discussed at present, but also for those above and below, in fact for all layers between 100 and 1000 metres (see also southern part of section I, salinity, fig. 16 A).

An area with a comparatively low temperature and a low salinity NE of the Talaud Islands indicates the region of divergence on the boundary between the North Equatorial (Mindanao) Current and the Equatorial Counter Current. Colder and less saline water rises to higher levels on this boundary.

400 m (figs. 21 D and 22 D).

The salinity at 400 m is in most places lower than at 200 m; this is in the first place due to the greater distance of the 400 m layer from the salinity maximum. Moreover, for the area near Mindanao, 400 m is approximately the depth of the salinity minimum of the north Pacific water. As

a result the difference in salinity between the Celebes Sea-Makassar Strait-Flores Sea water and the Halmahera Sea-Ceram Sea water is even greater than at 200 m. At the same time there is still a small difference in temperature between the two water masses, so that the distribution of Flores Sea water as against the Ceram Sea water in the Banda Sea can again be traced both by its lower salinity and its lower temperature (isohaline of 34.55 ‰ and isotherm of 9° C).

Much of the water present in the western Ceram Sea (Buru Sea) enters this area from the Halmahera Sea (compare p. 17). The area of high salinity north of Buru, however, is separated from the Halmahera Sea water by an area with a slightly lower salinity. The same isolated patch also occurs in deeper layers. In these layers no transport from the Halmahera Sea is possible, since the connection with this sea is cut off. Hence the increase of salinity, which is accompanied by a slight rise of temperature, must be due to the admixture of more saline water from higher levels.

The connection between the Celebes Sea and the Sulu Sea is interrupted at about 250 m. The Sulu Sea water at 400 m has a considerably higher temperature than the Celebes Sea water at the same depth, since the threshold between the Sulu Sea and the China Sea lies at about 400 m; vertical mixing over this sill, causing downward heat transport and a downward movement of the water layers immediately above sill level (VAN RIEL, 1956) explains the fact that inside the Sulu Sea the temperature at 400 m is higher than in the China Sea and therefore also higher than in the Pacific Ocean and the Celebes Sea.

600 m (figs. 21 E and 22 E).

Near Mindanao the salinity values are somewhat higher than at 400 m, because of the greater distance of the 600 m layer from the north Pacific salinity minimum. Near New Guinea, on the other hand, the salinities have decreased, because of the nearness of the salinity minimum at 800 m. Therefore, the horizontal differences in salinity at 600 m are comparatively small, although they are still sufficiently large (34.55 against 34.65 ‰) to allow a distinction to be made in the Banda Sea between the water entering this sea from the west and from the north. The two water masses, moreover, still show a small difference in temperature (6.5 against 7.5° C).

The connection between the Halmahera Sea and the Ceram Sea is cut off at about the depth under discussion, so that the only remaining connection between the Ceram Sea and the Pacific Ocean is through the Molucca Sea. The somewhat elevated salinity and temperature of the Ceram Sea and the areas around it must be due to vertical mixing, especially with more saline and warmer water from higher levels.

800 m (figs. 21 F and 22 F).

The passage between the Makassar Strait and the Flores Sea is closed above this depth, so that the only remaining open connection between the Pacific and the Indian Ocean is through the Molucca Sea and the Strait of Lifomatola. Horizontal differences in salinity and temperature in the Pacific Ocean water have practically disappeared. However, the mixing processes in the Ceram Sea area cause a somewhat higher salinity and temperature in this region.

The Banda Sea, which at this depth must receive practically all its water from the north, contains a very homogeneous water mass, which nevertheless still has a somewhat lower salinity and temperature in the south. Some water from the Strait of Makassar may therefore, although the connection is cut off, still influence the conditions in the western and southern part of the Banda Sea. Another possibility is, that the Flores Basin, which forms a somewhat isolated corner of the area investigated, contains older water from a period during which the salinity of the basins was lower than at the time of the expedition. The fact that the Flores Basin, and also the Sawu Basin, has a somewhat lower salinity than the Banda Basin down to the bottom is in favour of this possibility (see vertical sections IV and V, figs. 16, D and E).

The Halmahera Sea is at 800 m isolated in the north as well as in the south and both salinity and temperature have the values of the water at the threshold depth of 700 m.

1000 m (figs. 21 G and 22 G).

The salinity distribution at this level resembles the distribution at 800 m. The area with slightly elevated salinity values is still present in and around the Ceram Sea and in the eastern Banda Sea. Comparatively low salinities, down to 34.55 ‰, still occur in the Flores Sea, the western Banda Sea and the Sawu Sea.

The horizontal differences in temperature at 1000 m are very small and require no comment.

1250 m (figs. 24 A and 24 B).

At this depth the Sawu Sea is cut off from the Indian Ocean, but this does not influence the salinity and the temperature distribution in this area, since also in higher levels it was filled from the side of the Banda Sea. The only remaining communications between the Pacific Ocean and the Indian Ocean lie east of Timor and in the Ceram Sea. The horizontal distribution of salinity and temperature is very similar to the distribution at 1000 m and requires no further remarks.

B. *Water movement and renewal in the basins.* Van Riel (1934, 1943, 1956), by comparing potential temperatures and in some cases oxygen and salinity values at different depths on both sides of the thresholds separating the basins, arrived at the conclusion that by far the greater number of basins are filled from the Pacific side; only the Timor Trough and the Aru Basin are supplied from the Indian Ocean. The paramount influence of the Pacific Ocean is simply caused by the bottom configuration of the area: a basin is always supplied over the lowest sill and, with the two exceptions mentioned above, the deepest thresholds are on the side of the Pacific Ocean. Tydeman (1922) already arrived at the same conclusion.

Table 7

Principal basins: for location see fig. 3; *mainly according to* Van Riel (1934).

Name	Saddle depth, m	Maximum depth, m	Potential temp. of deep water	Salinity of deep water
Sulu Basin	400	5580	9.90	34.49
Mindanao Trench .	3500?	10500	1.20	.63
Talaud Trough . .	3130	3450	—	—
Sangihe Trough . .	2050	3820	2.15	$.64^{5}$
Celebes Basin . . .	1400	6220	3.30	.58
Morotai Basin . . .	2340	3890	1.55	.65
Ternate Trough . .	2710	3450	1.65	.67
Batjan Basin . . .	2550	4810	1.80	.66
Mangole Basin . . .	2710	3510	—	—
Gorontalo Basin . .	2700	4180	1.96	.63,
Makassar Trough. .	2300	2540	3.45	.58
Halmahera Basin .	700	2039	7.50	.60
Buru Basin	1880	5319	2.70	.61
N. Banda Basin . .	3130	5800	2.73	.62
S. Banda Basin . .	3130	5400	2.75	.60
Weber Deep	3130	7440	2.78	.61
Manipa Basin . . .	3100	4360	2.85	$.60^{5}$
Ambalau Basin . .	3130	5330	2.76	.61
Aru Basin.	1480	3680	3.70	.65
Butung Trough . .	3130	4180	—	—
Salajar Trough . .	1350	3370	3.67	$.60^{5}$
Flores Basin. . . .	2450	5130	2.98	.61
Bali Basin.	—	1590	3.50	.61
Sawu Basin	2100	3470	3.15	.61
Wetar Basin. . . .	2400	3460	2.95	.61
Timor Trench . . .	1940	3310	2.55	.71
Sunda Trench . . .	—	7140	0.77	.71

These facts are well illustrated by considering the temperature distribution of the vertical sections (figs. 17, A-G). Moreover a table published by VAN RIEL, containing the values for saddle depths, maximum depths, temperature and salinity characteristics of the various basins is republished here for convenience of the reader, after some slight modifications.

The basins are filled with water masses in which the differences in potential temperature never exceed a few tenths of a degree Celsius. Since in the oceans the temperature decreases with depth, the temperature in a basin will be lower when the saddle depth is greater. The upper boundary of the true "homothermal" deep, however, always lies a few hundred metres below the depth of the threshold; this must be due to downward heat transport by vertical turbulence in the water passing over the sill and by the downward movement of the water layers that lie immediately above sill level. Consequently, of two basins, one located behind the other, the second basin will have warmer bottom water than the first, even when the sill between the two basins is lower than the sill between the first basin and the ocean (GROEN in *Van Riel*, 1956); when the second sill is higher than the first, it goes without saying that temperatures in the second basin are higher than in the first. Hence, in a series of adjoining basins a basin is always filled with warmer water than its predecessor, independent of the depth of the separating threshold; examples will be given below.

Also the salinity of the basins depends on the sill depth, but in this case the relation is not so simple, since in the open oceans the vertical distribution of salinity is more complicated than the distribution of temperature. In the Pacific Ocean the salinity first increases with depth to about 150 m, then decreases to about 400 m or 800 m, depending on the region considered, and increases again below these depths. In the Indian Ocean similar conditions exist.

If we consider the temperature and salinity distribution of the sections in more detail, we find in Section I (figs. 16 A and 17 A), that the Pacific water first enters the Sangihe Trough at a depth of 2050 m, filling this basin with water of a potential temperature of 2.15° C. The Celebes Basin with a saddle depth of 1400 m, north of the Talaud islands, receives water with a comparatively high potential temperature of 3.3° C and a rather low salinity of 34.58 ‰. The Makassar Trough obtains water from the Celebes Sea over a sill at a depth of 2300 m with a potential temperature of 3.45° C. In the south this trough is closed off and separated from the Flores Basin by a shallow sill of 650 m depth, over which no deep water passes.

A much longer series of basins is traversed by section II (figs. 16 B and 17 B). First, Pacific water enters a number of small basins in the Molucca Sea with very deep sills (2340-2710 m). These basins therefore contain comparatively cold water (1.55-1.80° C) with the fairly high salinity of 34.65 ‰. West of the island of Obimajor the water passes over a sill of 1880 m into the Buru Basin. This passage (Lifomatola Strait) is of the greatest importance, since it forms the only intake for the deep water of a large complex of basins, including the Buru Basin, the northern and southern Banda Basin, the Flores and Bali Basin, the Sawu Basin, the Aru Basin and a few smaller basins, all listed in table 7. The deepest connection between this system of basins and the Timor Trench is located east of the island of Timor and attains a depth of about 1400 m.

The Buru Basin, the first basin behind the Strait of Lifomatola, obtains water of a potential temperature of about 2.7° C and a salinity of 34.61 ‰. NW of the island of Buru water passes over a deep sill of 3130 m into the Banda Basins (N. and S. Banda Basin, Weber Deep, Manipa Basin and Ambalau Basin), which receive water of a potential temperature between 2.70 and 2.85° C and the same salinity as the water of the Buru Basin.

Section IV (figs. 16 D and 17 D) follows a traverse through the Flores and the Banda Basins from west to east and crosses Section II at station 246. The Flores Basin is filled from the east over a sill of 2450 m with water of which the potential temperature is somewhat below 3° C. In the "homothermal" deep of the Flores Basin isotherms have been drawn representing intervals of only 0.01° C. The regularity of these isotherms gives a very favourable idea of the accuracy of the temperature measurements. The Bali Basin forms an appendix to the Flores Basin without a saddle in between. On its eastern side section IV crosses the Aru Basin, which is not filled, however, from the side of the Banda Sea, but from the Indian Ocean via the Timor Trench (VAN RIEL, 1943).

The small Wetar Basin and the Sawu Basin are traversed by section V. The threshold between the former basin and the Banda Sea lies at 2400 m; the potential temperature of the water is about

2.95° C and the salinity is the same as that of the Banda Sea (34.61 $^0/_{00}$). The Sawu Basin has a saddle depth of 2100 m, a potential temperature of 3.15° C and again a salinity of 34.61 $^0/_{00}$. It is separated from the Indian Ocean by a threshold of about 1100 m in the Sawu Strait.

An exceptional position is occupied by the Halmahera and Sulu Basins. Both are connected with the ocean by very high sills. The Halmahera Basin is directly connected with the Pacific Ocean over a threshold of about 700 m between the island of Halmahera and New Guinea. The bottom temperature of the basin is accordingly high (7.8° C) but its salinity is rather low (34.60 $^0/_{00}$). In the south the Halmahera Basin is separated from the Buru Basin by a sill of about 500 m.

The Sulu Basin is filled from the north with Pacific water which enters the area via the China Sea. The sill depth is about 400 m and the temperature of the deep water in the basin about 10° C. The salinity is comparatively low (34.49 $^0/_{00}$). The deepest connection between the Sulu Sea and the Celebes Sea (Sibutu Strait) has a depth of only 260 m.

As regards the basins filled from the Indian Ocean, it is worthy of note that the deep water of the Indian Ocean attains a lower temperature than that of the westernmost part of the Pacific Ocean (0.8° C against 1.3° C). This difference is mainly caused by the Philippine Ridge between New Guinea and Japan, which below 3500 m obstructs the passage of cold Antarctic deep and bottom water from the central Pacific Ocean to the west. In the Indian Ocean no such obstruction exists.

The Timor Trench is separated from the Indian Ocean by a threshold of 1940 m depth; the water flowing over the sill has a potential temperature of 2.55° C and a salinity of 34.71 $^0/_{00}$. The Aru Basin is separated from the Timor Trench by a saddle depth of 1480 m filled with water of a potential temperature of about 3.7° C and a salinity of 34.65 $^0/_{00}$. On its north side the Aru Basin is separated from the Buru Basin by an area where the depth is not much less than 1500 m. Since, therefore, the sills north and south of the Aru Basin have approximately the same depth, there is from the point of view of bottom topography no decisive reason why no water should penetrate into the Aru Basin from the north. Also the temperature and the oxygen distribution are not conclusive in this respect. Van Riel's argument for a supply from the south is that the salinity of the Aru Basin (34.65 $^0/_{00}$) corresponds with the salinity at a depth of about 1500 m in the Timor Trench and is higher than in the Buru Basin at the same depth (34.61 $^0/_{00}$). Considering the salinity distribution along section III, it seems more probable to the present author, however, that the Aru Basin contains a mixture of water from the north and the south.

Besides the direction of water supply, a second important factor to be considered in connection with the chemical investigations is the *rate of water renewal* of the basins. Already during the investigations of the "Siboga"-expedition (1900) attention was drawn to the presence of very coarse sedimentary material on the floor of the passages from one basin to another. These coarse sediments indicate the existence of rapid currents over the sills and Tydeman (1922) already assumed that large amounts of water must therefore be supplied to the basins.

The existence of strong currents over the thresholds was proved by the current measurements of the "Snellius"-expedition (Lek, 1943). Of special importance in this connection are the current observations in the Lifomatola Strait; here, at a depth of 1500 m, tidal currents attain velocities up to 30 cm/sec. These currents will be responsible for the coarseness of the sediments. Moreover, the observations prove the existence of a residual current of about 5 cm/sec through this strait into the Buru Basin.

The constant supply of new water to the basins requires an explanation: at first sight one would expect that, once a basin is wholly filled with water of the appropriate density, no further supply would be needed and the inflow of water would come to an end. Many small and large basins are known, where such stagnant conditions occur with all their chemical consequences, such as depletion of oxygen, increase of nutrient concentrations and, in many cases, formation of hydrogen sulphide. Evidently, in the Indonesian basins (with the exception of the Kau Bay), stagnancy does not occur. In fact, as we shall see in chapter IV, water renewal is so fast that only a slight consumption of oxygen takes place inside the basins.

Tydeman (1922) supposes that this constant renewal of water is caused by heat flow from the earth's crust. He reasons that this flow must increase the temperature of the bottom water inside the basins. Convective water movements will pass the heat towards higher levels, so that the whole homothermal mass is slowly warmed up. Outside the basins, in the open ocean, such an increase in

temperature also occurs, but here the supply of cold deep water from polar regions keeps the temperature at a constant level. As a result a difference in temperature, i.e. a difference of specific weight, is brought about between the basin and the open ocean, so that the situation on either side of the sill becomes unstable. Cold water from the ocean enters the basin over the sill and the difference in temperature is kept at a constant level.

KUENEN (1943) continued this line of thought by assuming that the ventilation of the basins is caused by the combined action of heat flow from the earth's crust and tidal currents. Both effects will lead to an increase in temperature of the deep water, the former from below and the latter from above, owing to the mixing of water from different levels.

Vertical mixing certainly plays a role in the layers at threshold depths, as is proved by the depression of the isotherms at these depths, discussed previously. It may also be of importance in deeper water of the basins, where the tidal currents are often surprisingly strong and differences in density very small. An estimate of its relative importance for the renewal of the water seems difficult, however; KUENEN advances arguments that ventilation by thermal convection is at least of equal importance as renewal by tidal currents. We will return to this problem on p. 54.

The same author has tried to determine the maximum time needed for complete water renewal of the Banda Basins by assuming that this renewal is caused by heat transport from the earth alone. For this calculation he supposes that the amount of heat flow under the basins equals that of the continents and that water should be heated 0.1° C before it will leave the basin. The first assumption seems to be justified by recent measurements of heat transport through the ocean bottom (BULLARD e.o., 1954, 1956). The second assumption may be rather arbitrary, since in fact the difference in temperature between the water leaving and entering the basin is unknown. The result of the calculation, however,—a renewal of the "homothermal" water mass in 200-300 years at most—tallies very well with a calculation made along an entirely different way. This calculation was also first worked out by KUENEN (1943), but a slight numerical correction of his result seems necessary.

The thickness of the current passing southward through the Strait of Lifomatola may be estimated at 400 m. The cross-section of this strait between the threshold at 1880 m and 1480 m is about 4 km^2 (VAN RIEL, 1956). The current has a southward component of 5 cm/sec at 1500 m (LEK, 1953; Kuenen assumed 7 cm/sec); taking this value as an average for the whole cross-section, the yearly intake into the basins is 6300 km^3. The combined surface of all basins supplied by the Lifomatola Strait is about 900.000 km^2 and the depth of the "homothermal" mass some 2 km. Thus a complete renewal of the homothermal waters in these basins will take place in about 300 years.

Taking all things together, a period of about 300 years may be a fairly good maximum estimate. The reliability of this value will be discussed in chapter IV. The question may already be raised here, however, in how far one series of current determinations suffices as an estimate of conditions for a long period. Even without questioning the current measurements as such, one might suppose that the estimate is only valid for a rather short period, because the rate of flow through the inlet decreases and increases periodically or because this flow is irregular. If this were true, only frequently repeated observations could yield a reliable estimate of water transport.

In this connection it may be observed that recent observations and discussions concerning the deep water circulation of the North Atlantic Ocean have revealed that the renewal of this water mass occurs irregularly (WORTHINGTON, 1954, COOPER, 1956, DIETRICH, 1956). The causes for these irregularities are probably changes in the amount of deep water available at the source, brought about by atmospheric changes, such as more or less severe winters, combined with changes in the wind and barometric pressure field over the ocean, which sometimes promotes, but in other periods prevents the flow of water over the sills between England and Greenland.

In the tropical Indonesian area climatic irregularities will probably be of less significance than in the temperature and polar regions of the North Atlantic Ocean. Moreover, the deep water filling the basins is a very long distance from its source, so that irregularities in its supply may be partly or completely smoothed out. Changes in the rate of flow due to seasonal influences seem quite possible, however. Only continued research can provide a satisfying answer in this respect.

C. *Temperature-salinity relations.* A valuable means to obtain a general view of the relations between the different water masses is provided by the TS-diagrams. For every more or less uniform

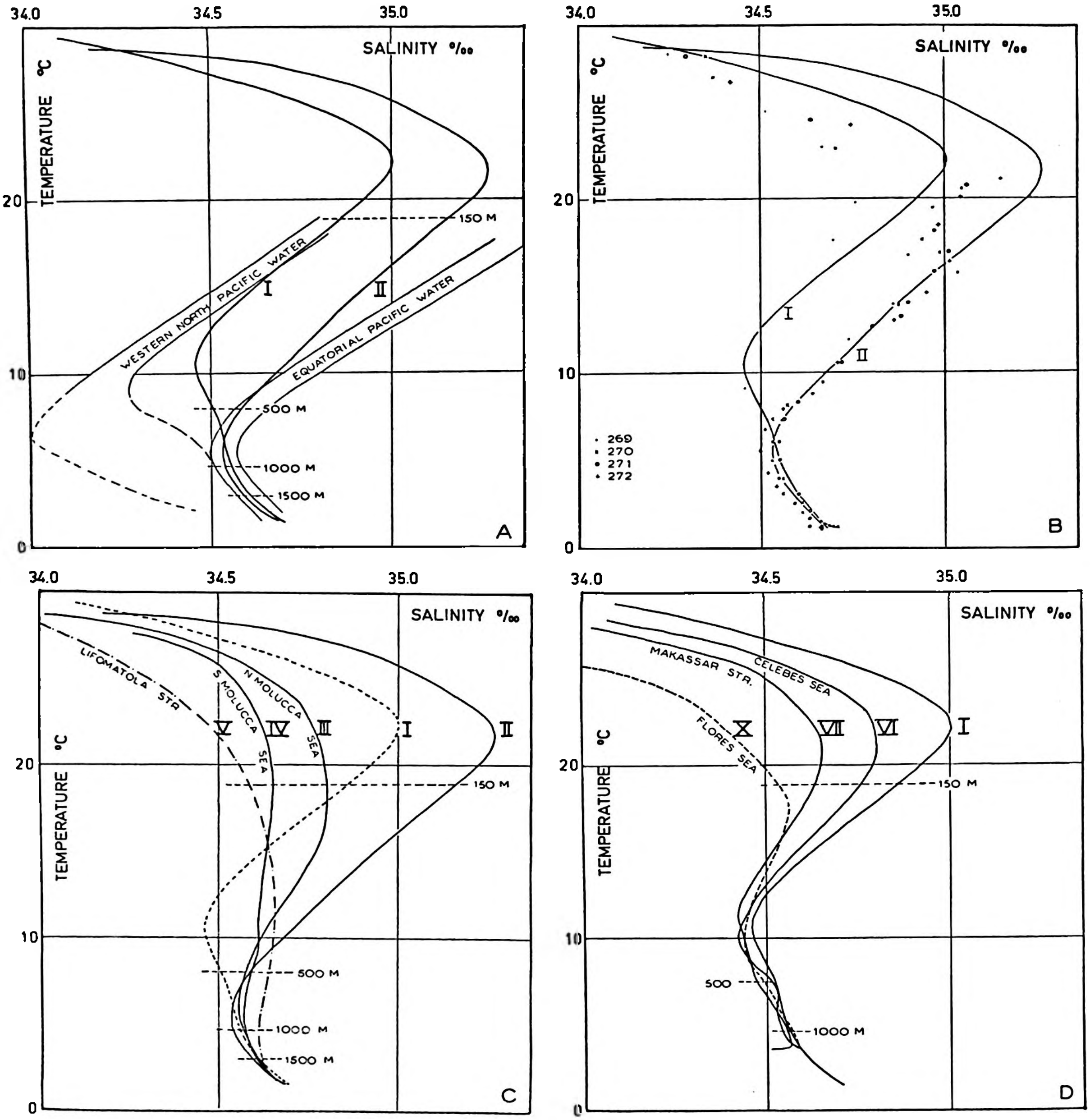
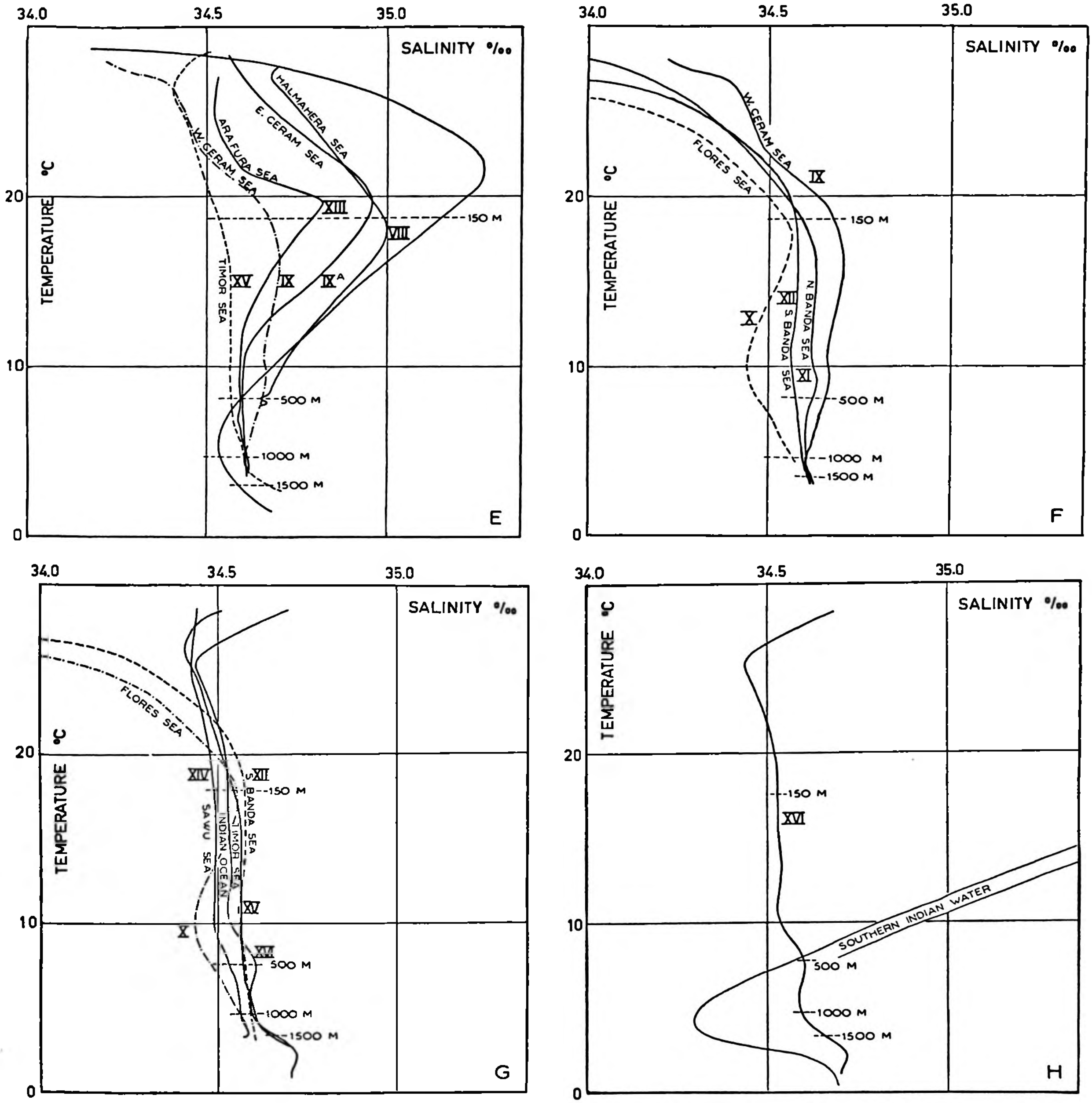


Fig. 7. Temperature-salinity relations; compare table 8.

area a number of stations has been combined to obtain the average TS-curve of this area (table 8). The TS-relations of the various areas are in their turn taken together in a series of graphs (figs. 7 A-7 H).

The first figure (7 A) represents the "Snellius"-observations in the Pacific Ocean between



Mindanao and New Guinea (curves I and II), together with the TS relations of the two water masses in the Western Pacific Ocean—the *Western North Pacific Central* and *Equatorial Pacific* water, which have been taken from SVERDRUP et al. (1946). The former water mass includes the water of the North Equatorial Current, the latter that of the South Equatorial Current. Curve I represents

the conditions near Mindanao, curve II those near New Guinea. Below about 6° C (700-800 m) both areas follow the relation of the Equatorial water; above 800 m the influence of the North Equatorial Current causes curve I to approach the curve of the WN Pacific water.

The transition between the two water masses lies in the area east of the Talaud islands (stations 269, 270, 271 and 272, fig. 7 B) in the Molucca Sea (fig. 7 C).

Table 8

TS-diagrams; compare figs. 7, A-H.

	Area	Stations
I	Pacific Ocean, North	262-268, 293-296
II	Pacific Ocean, Equatorial	273-276
III	Molucca Sea, North	283-288, 290-292
IV	Molucca Sea, South	80, 332-336
V	Lifomatola Strait	223-228
VI	Celebes Sea	47, 48, 52, 53, 56, 57, 75, 76, 301, 302, 303, 304
VII	Makassar Strait	35, 39, 41, 310, 311, 312
VIII	Halmahera Sea	353
IX	Ceram Sea, West (Buru Sea)	229, 330
IX^A	Ceram Sea, East	87-91
X	Flores Sea	167, 179, 180, 197
XI	Banda Sea, North	209, 210, 212, 215, 218, 331
XII	Banda Sea, South	205, 235, 238, 240, 241, 245, 246, 249, 319
XIII	Arafura Sea	100, 104
XIV	Sawu Sea	155, 159, 160, 163, 378, 379, 380
XV	Timor Sea	115, 118, 121, 125, 127
XVI	Indian Ocean	145-147, 382

The most important feature of the transition area is the low salinity of the salinity maximum, the cause of which has already been discussed on p. 16. It must immediately be added here that judging from its oxygen content the Molucca Sea only contains North Equatorial water.

For the Molucca Sea (fig. 7 C) separate curves have been drawn representing its northern part (III) and its southern part (IV), while also the curve for Lifomatola Strait (V) and the two curves for the Pacific Ocean are given in the same figure. The lower parts of the Molucca Sea curves clearly represent an intermediate form of the curves I and II. It is not impossible that part of the WN Pacific Central water does not directly enter from the Pacific Ocean, but has made a detour through the Celebes Sea. In the Strait of Lifomatola the influence of less saline water near the surface causes the upper part of this curve to bend far towards the left. Below about 12° C (250 m) the salinity in this strait is higher than in the Molucca Sea, probably partly owing to the influence of saline water from the Halmahera Sea (between 250 and 500 m) and partly as a result of vertical mixing in the narrow strait, which straightens the lower part of the TS-curve of the southern Molucca Sea.

The condition in the water masses flowing from the Pacific Ocean into the Celebes Sea (VI) and from there into the Makassar Strait (VII) and the Flores Sea (X) are represented in fig. 7 D. The WN Pacific water, as represented by curve I, is during this transport gradually transformed owing to vertical mixing, together with admixture of less saline water from above. However, the salinity minimum changes very little and is still distinctly present in the Flores Sea.

The TS-diagrams for the water masses in the seas west of New Guinea are given in fig. 7 E, together with the curve for the Pacific Ocean near New Guinea (II). Already in the Halmahera Sea the salinity maximum is reduced, probably as a result of vertical mixing, and for the same reason

the salinity below 12° C is higher than that of curve II; it must be taken into account that the threshold between the Halmahera Sea and the Pacific Ocean attains a depth of only 700 m.

As a result of this increase in salinity, together with the increase in Lifomatola Strait referred to above, the deeper layers of the western Ceram Sea (IX) contain water with a comparatively high salinity. In view of this fact the salinities of the intermediate water in the eastern Ceram Sea (IX^A) and the Arafura Sea are surprisingly low. This can only be explained from the fact, that a large part of these water masses do not come from the north, but from the west out of the Banda Sea (fig. 7 F) and the Timor Sea (XV). Below about 700 m (6.5° C) the direct connection between the Arafura Sea (Aru Basin)-East Ceram Sea and the Banda Sea is closed and the Banda Sea water has to be supplied via the Timor Sea. Below 1000 m the differences in salinity between the Timor Sea and the western Ceram Sea seem too small for us to decide whether the water of the Arafura Sea is supplied from one area or the other (compare p. 22).

The TS-relations for the Banda Sea, represented by a curve for the northern (XI) and another for the southern part (XII) of this area, are given in fig. 7 F, together with the curves for the Flores Sea and the western Ceram Sea. The two Banda Sea curves can be considered as resulting from a mixture of water from these two areas with the greater influence of the Ceram Sea in the north and of the Flores Sea in the south. The slight bend to the right in the western Ceram Sea curve between 400 and 500 m, caused by the inflow of Halmahera Sea water in this area, is also present in the northern Banda Sea relation. The salinity minimum of the Flores Sea can still be weakly discerned in the southern Banda Sea.

Fig. 7 G gives the curve of the "Snellius" observations in the Indian Ocean (XVI), together with the curves of the adjacent Indonesian seas. Fig. 7 H represents the Indian Ocean curve together with the TS-relation of the Southern Indian Ocean Water according to SVERDRUP et al. (1946). According to LOTTE MÖLLER (1929, 1933) this water mass prevails in the area south of Indonesia.

Above about 400 m (9° C) curve XVI deviates so much from the relation for the Indian Ocean, that the water between this depth and the surface layer can only originate from the Indonesian seas. This point of view is confirmed by the fact that in fig. 7 G curve XVI from 400 m to about 150 m lies between the Sawu Sea (XVI) and the Timor Sea (XV) curves and therefore consists of a mixture of water from these two seas.

Between 150 m and the surface the origin of the water depends on the season.

Below about 400 m the curve XVI lies to the right of the curves for the southern Indonesian seas; it must, therefore, at least contain a certain amount of Indian Ocean water. The importance of this admixture cannot be estimated from the TS-curves alone; it will be shown in chapter IV that the Indian Ocean influence is small down to a depth of about 1000 m and then increases.

The part of the Sawu Sea curve between 500 and 1100 m (the threshold depth of the southern entrance of the Sawu Sea) may be interpreted in two ways. It can be considered as consisting of a mixture of Flores Sea and S. Banda Sea water, but also as a mixture of Flores Sea water and the water represented by curve XVI. On previous pages we already saw, that the current measurements indicate a penetration of water from the south into the Sawu Sea between 400 and 800 m; in favour of the latter supposition is also the fact, that the Sawu Sea curve at these depths follows the slight bend to the right of curve XVI. We shall return to this problem in connection with the distribution of oxygen.

D. *Summary of water movement.* A short summary may finally be given of the main conclusions arrived at in the present chapter. The most conspicious feature of water movement is the predominance of the Pacific Ocean over the Indian Ocean, not only below the thresholds, where this phenomenon was already described by VAN RIEL, but also between the sill levels and the surface. A fortunate circumstance in determining the direction of flow is the fact that the water masses in the adjacent part of the Pacific Ocean possess a very distinct structure, the origin of which could be traced without difficulty. Towards the southern part of the Indonesian Archipelago this structure becomes gradually less conspicuous and in the Banda Sea and the adjacent part of the Indian Ocean a comparatively homogeneous water mass is encountered. There can be little doubt that the direction in which the distinction between the water masses becomes less clear corresponds with the main direction of flow. This conclusion is supported, moreover, by the direct evidence of the current measurements. (LEK, 1938).

The intermixture of water masses in the East Indonesian seas will to a considerable degree be promoted by their passage through narrow straits, where strong currents occur, and by the tidal streams, which are often very strong even at great depths (LEK, 1938) and certainly play a much more important rôle here than in the open oceans.

The main cause of the Pacific influence in the deep water is the greater depth of the thresholds on that side. A factor inducing the transport towards the Indian Ocean in the upper water layers may be the piling up of water against the western boundary of the equatorial Pacific Ocean by the prevailing winds. Below these layers and above the thresholds the transport towards the Indian Ocean must be controlled by the field of density, but a discussion of this aspect falls outside the scope of this study. Near the surface the direction of water movement is to a large extent dependent on the wind field and therefore changes with the seasons.

In the part of the Pacific Ocean adjacent to the East Indonesian seas a clear distinction can be made between the water masses of the North and the South Equatorial Current. The North Equatorial Current has a lower maximum salinity and, as will be shown in chapter IV, a higher oxygen content than the South Equatorial Current. The boundary between the two water masses, which shifts with the seasons, lies in the period from June to November (when the South Equatorial Current has its greatest westward extension) in the area east of the Talaud Islands and north of Halmahera.

The region between the two currents is further marked by low salinity or the absence of the salinity maximum. This condition is characteristic of the Equatorial Counter Current and from this point of view also the Molucca Sea can be considered as belonging to the transitional region of the Counter Current.

The lower limit of the equatorial currents lies at a depth of about 150 m, and below this level we find in the Pacific Ocean two intermediate water masses, one of North Pacific and another of South Pacific origin (North and South Pacific Intermediate Water), both of which are characterized by a salinity minimum. The minimum of the first water mass has a lower salinity (34.45 ‰) and lies at a smaller depth (400 m) than that of the second water mass (34.55 ‰ and 800 m). There is no distinct boundary between these two water masses; the intermediate water south of the Talaud Islands is chiefly of southern origin, that north of these islands is above 800 m a mixture of southern and northern water; below 800 m all water between Mindanao and New Guinea is South Pacific water.

As a result of these differences between the water masses of the Pacific Ocean from north to south, the water flowing into the Celebes Sea has a structure different from that entering the Molucca Sea and this water in its turn differs from the water penetrating into the Halmahera Sea. For the study of the water movements in the Indonesian seas the existence of these differences is very fortunate, since it enables us to distinguish between the various water masses down to the extreme south of the Archipelago.

The Celebes Sea water can in this manner be followed during its transport through the Strait of Makassar, the Flores Sea, the Sawu Sea and the southern Banda Sea; its influence can also be discerned in the Arafura Sea, the Timor Sea and the Indian Ocean. Taking all things together, the flow of water around the western side of Celebes is by far the most important water transport quantitatively. The Molucca Sea water is found back in the western Ceram Sea, the northern Banda Sea and also in the southern Banda Sea. The Halmahera Sea water is found in the latter areas and also in the Arafura Sea.

The details of the various water movements have been discussed extensively in the previous pages and will not again be entered upon. We may stress, however, that the influence of the earth's rotation, although small at these low latitudes and, moreover, obscured by the complicated configurations of many areas, must nevertheless exert a certain influence on the direction of the various water movements. How far, however, these movements are controlled by the Coriolis force can only be ascertained by hydrodynamical considerations (see Vol. II, Part 9, which will appear later).

Attention is finally drawn to the comparatively rapid renewal of the deep water in the basins. According to KUENEN the water in the homothermal layers of the basins filled through the Strait of Lifomatola are renewed once in about 200-300 years. One important factor causing the water renewal may be assumed to be the flow of heat through the sea bottom from the earth's crust: another is the warming of the deep water by mixing from above.

CHAPTER FOUR

CHEMICAL INVESTIGATIONS

In the following pages the four chemical properties determined besides salinity will be discussed successively. Oxygen will be dealt with first. The hydrogen ion and alkalinity data will be examined in conjunction, since both are connected with the same problem, that of the precipitation and solution of lime. Finally, the phosphate observations will be considered.

The main sections, used in the previous chapter to elucidate the vertical distribution of salinity and temperature (fig. 2), are also employed to represent the distribution of the chemical properties. Horizontal sections, in depths corresponding with those for salinity and temperature, have been constructed for oxygen only; in the other cases such charts did not bring new distribution characteristics to the fore, so that their reproduction seemed unnecessary. For phosphate, the number of observations available is too small to allow the construction of cross-sections.

A. *Oxygen.* 1. *Vertical sections.* The characteristics of the oxygen distribution in a given area must be explained on the basis of both water movements and biological processes. Below the euphotic zône the latter always lead to consumption of oxygen. The water movements must therefore cause a replenishment of oxygen, which in the state of equilibrium balances the consumption. One of the principal aims of the present discussion is to determine the influence on the oxygen distribution in the Indonesian seas of both water movements and oxygen consumption in situ. In this connection, first the relation with water movements will be studied.

For a description of the distribution of oxygen, the vertical sections I-III (figs. 18 A-G, pp. 75-79) form the best starting-point.

The conditions in the Pacific Ocean are clearly elucidated by the oxygen distribution in section VII (fig. 18 G). North of the Talaud Islands an oxygen minimum is present at a depth of about 400 m; a second minimum is present both north and south of these islands at a depth of about 1000 m. The first minimum is also present in section I (fig. 18 A) and is found to have an irregular appearance. The second minimum occurs in the sections I, II and III (figs. 18, A-C).

Both minima clearly originate from the Pacific Ocean, which is in accordance with the source of the corresponding water masses. The oxygen distribution in this ocean is characterized by a minimum with a very low oxygen content in its northern and central part. Near the American coast an area even completely devoid of oxygen is found (Sverdrup et al., 1946), but also farther to the west the oxygen content is lower than 1 ml/l over large areas.

In the northern Pacific the oxygen minimum layer lies at a depth of about 1000 m, but near the Equator it rises to a depth of 400 m. Its southern boundary lies near the Equator. This boundary has been determined by several expeditions, among which the "Carnegie", the "Dana", the "Albatross' and the "Hugh M. Smith" expeditions. The westernmost oxygen section available is that of the "Planet" (1906-1907) between eastern New Guinea and the Philippines. In this area the southern limit of the oxygen minimum layer, judging from the isopleth of 2 ml/l, is located at 3° N, also at a depth of about 400 m.

South of the Equator the oxygen values are never as low as in the North Pacific Ocean. In the Antarctic Ocean an oxygen minimum develops between the Antarctic Circumpolar Water and the Antarctic Intermediate Water ("Discovery" section in Sverdrup et al., 1946, p. 622). Since these two water masses slope down towards the north the depth of the oxygen minimum also increases and at a latitude of 40° S it lies at a depth of about 2000 m. At the same time the oxygen content of the minimum decreases, but never falls below 3.0 ml/l. This minimum gradually disappears to the north, owing to the more rapid decrease of oxygen in the surrounding water masses. Especially

in the Antarctic Intermediate water the oxygen decrease is faster, so that near the Equator the southern oxygen minimum is located at a depth of about 1000 m, a few hundred metres below the centre of the latter water mass ("Dana" section in SVERDRUP et al., 1946, p. 728).

The oxygen minimum at 400 m of the "Snellius"-section I evidently corresponds with the northern oxygen minimum, whereas the minimum at 1000 m of the sections I, II and III represents the southern oxygen minimum. The two oxygen minima closely correspond with the respective salinity minima, although they do not exactly coincide.

The irregularities of the northern oxygen minimum observed in the sections I and VII, just as the irregularities in the northern salinity minimum discussed in chapter III, indicate that the North Pacific Intermediate Water which enters the area near Mindanao, comes from somewhat different latitudes. Similar irregularities occur in the open Pacific Ocean; the "Hugh M. Smith" observations in the equatorial region south of the Hawaiian Islands show, for example, that in one period the northern oxygen minimum penetrates farther to the south than in another and that even isolated nuclei of oxygen occur (CROMWELL et al., 1951-1954).

The irregularities of the 400 m oxygen minimum in the Celebes Sea quickly vanish when the water is transported to the west. In section VI, which crosses the western Celebes Sea from north to south, and in Makassar Strait, no further traces of this minimum are found.

The oxygen minimum at 1000 m, which, as we already saw, is well-pronounced in the eastern part of section I, passes just above the threshold of the Sangihe Trough and the Celebes Basin. Hence the water mass flowing into the latter basin is derived precisely from the layer where this minimum is found. This fact explains the comparatively low oxygen content of the Celebes Basin.

The same oxygen minimum layer also penetrates into the Molucca Sea. Towards the south the minimum becomes rapidly less pronounced. It thus undergoes the same modification as the corresponding salinity minimum (p. 17). As a result only very small vertical differences occur in the water entering the Ceram Sea, and from there into the Banda Sea, through the Strait of Lifomatola.

Part of the water flowing into the Ceram Sea, however, originates from the Halmahera Sea; this holds especially for the water layers between 250 and 600 m (p. 17). This transport not only causes a considerable increase of salinity at the stations 221, 229, 225 and 227 of section II, but also an increase of the oxygen content at these stations. To explain this increase, the oxygen distribution in the Halmahera Sea must be considered. The shallowness of the Halmahera Sea thresholds prevents the penetration of the oxygen minimum at 1000 m into this area (section III). Moreover, the oxygen minimum at 400 m is practically absent here. Hence the water passing through the Halmahera Sea has a comparatively high oxygen content.

Taking all things together, the water masses entering the northern Banda Sea have at all depths a comparatively high oxygen concentration, which never falls below 2.4 ml/l. It is surprising, therefore, that in the southern and western Banda Sea (section I, south of station 209, and section IV) the oxygen minimum is again established, although at a smaller depth (800 m) than that in the Molucca Sea. As the most obvious explanation one will be inclined to assume, that in the Indonesian Archipelago the oxygen minimum is formed anew by oxygen consumption in the water masses during their displacement to the south. However, two other factors must be considered, which are probably of more importance than oxygen consumption in situ.

First, it has been shown in the previous chapter that water of the northern Pacific, entering through the Celebes Sea and the Strait of Makassar, plays an important rôle in the Flores Sea, the southern Banda Sea, the Sawu Sea and the Arafura Sea. This water mass has a comparatively low oxygen content, which just above the sill between Makassar Strait and the Flores Sea amounts to 2.1—2.2 ml/l (section I). A comparison of the oxygen and the salinity distribution in the sections II, III and IV may elucidate the relation between these two factors; the oxygen minimum always lies a few hundred meters below the salinity minimum, but otherwise the two distributions are closely connected.

In the second place, in the previous chapter the possibility could not be excluded that in the layer below 400 m Indian Ocean water enters the southern Banda Sea, presumably through the Sawu Sea. It is therefore of importance to give attention to the oxygen distribution in this ocean.

The distribution in the Indian Ocean is also characterized by the presence of an oxygen minimum at intermediate depths. At latitude 50° S this minimum, just as in the southern Pacific Ocean,

lies at a depth of about 2000 m. Towards the north the depth decreases and near the Equator it has ascended to 600-800 m. At the same time the oxygen content decreases to 1.0—1,5 ml/l, while north of the Equator values below 0.5 ml/l are encountered. Moreover, in this region a second oxygen minimum has developed at a depth of 400-500 m. These conclusions are based on the observations of a number of expeditions (SVERDRUP et al. 1946, CLOWES and DEACON, 1935, MÖLLER, 1933).

South of the island of Java, according to observations of the "Dana" and "Albatross" expeditions, the oxygen minimum lies at about 600 m; the oxygen content increases from about 1.5 ml/l near western Java to 1.9 ml/l near eastern Java. The minimum in the southern Banda Sea can therefore be considered as an extension of the Indian Ocean minimum.

Taking all things together, the oxygen minimum in the southern part of the Indonesian Archipelago may owe its existence to three influences: (1) inflow of North Pacific Intermediate Water with a comparatively low oxygen content, (2) penetration into the Archipelago of the oxygen minimum of the Indian Ocean and (3) consumption of oxygen in situ. A choice between these possibilities can be made as follows.

Fig. 8A represents relations between salinity and oxygen content at a depth of 600 m. The centre of the oxygen minimum under consideration lies at this depth; moreover, the connection between the Flores Sea and the Strait of Makassar is just not interrupted. The salinity-oxygen relations have been drawn for the Banda Sea, including the western Ceram Sea, for the Flores Sea, the Sawu Sea and the Timor Sea and finally, for the part of the Indian Ocean near Indonesia. In order to complete the latter diagram, data have been added from stations of the "Dana" and "Albatross" expeditions south of Java. The conditions above the southern threshold of the Makassar Strait are represented by station nr. 185; they are representative for the whole southern part of this strait.

We find, in the first place, that the points of the Flores Sea and Sawu Sea are closely arranged around station 185; in the second place the salinity-oxygen relation of the Banda Sea is an approximately straight line between station 185 and the stations in the Ceram Sea; thirdly, in the Indian Ocean the oxygen content decreases with the salinity, while the lowest part of the salinity-oxygen line coincides with the Sawu sea relation: finally, the salinity-oxygen relation for the Timor Sea closely corresponds with that of the Sawu Sea, but the salinity is somewhat higher.

A natural explanation of these facts is afforded by the assumption that the oxygen content of the water masses is simply determined by the oxygen content of three sources: the Ceram Sea, the Strait of Makassar and the Indian Ocean South of Java. The Flores, Timor and Sawu Sea water masses, according to this point of view, contain nearly unmodified water from the Makassar Strait. The Banda Sea water is a mixture of Ceram Sea and Makassar Strait water. In the Indian Ocean the portion of Indonesian water in the mixture is important close to the island of Timor, but its importance decreases towards the west. The Timor Sea probably contains water with a somewhat higher salinity from near New Guinea.

It appears therefore that the oxygen minimum in the southern part of the Indonesian seas mainly owes its existence to the supply of Pacific water via the Celebes Sea and the Strait of Makassar. Actually, this water mass appears to dominate at a depth of 600 m and also in higher levels. This is certainly due to the fact, that the current velocities in the passages west of Celebes are much higher than in these on the east side of this island (LEK, 1938).

At a depth of 650 m, however, the connection between the Strait of Makassar and the Flores Sea is interrupted, so that below this depth no further supply from this direction is possible. If we take into account that the residual currents through the remaining connections with the Pacific Ocean are weaker than through those west of Celebes, the possibility must be considered that below 600 m Indian Ocean water easily penetrates into the Indonesian seas.

It would be comparatively simple to trace this influence, if we had a comprehensive knowledge of the distribution of salinity and oxygen in the adjacent part of the Indian Ocean. The data available are restricted, however, to a rather narrow zone south of Java, where a number of deep sea expeditions carried out their investigations. Below about 10° S no good observations are available.

It would be especially impoıtant to obtain a north-south section of salinity and oxygen west of Australia. At present it is only possible to construct, on the basis of the "Snellius"-data themselves, a short section from the Lesser Sunda Islands to the Australian shelf (fig. 8B), which shows the following details.

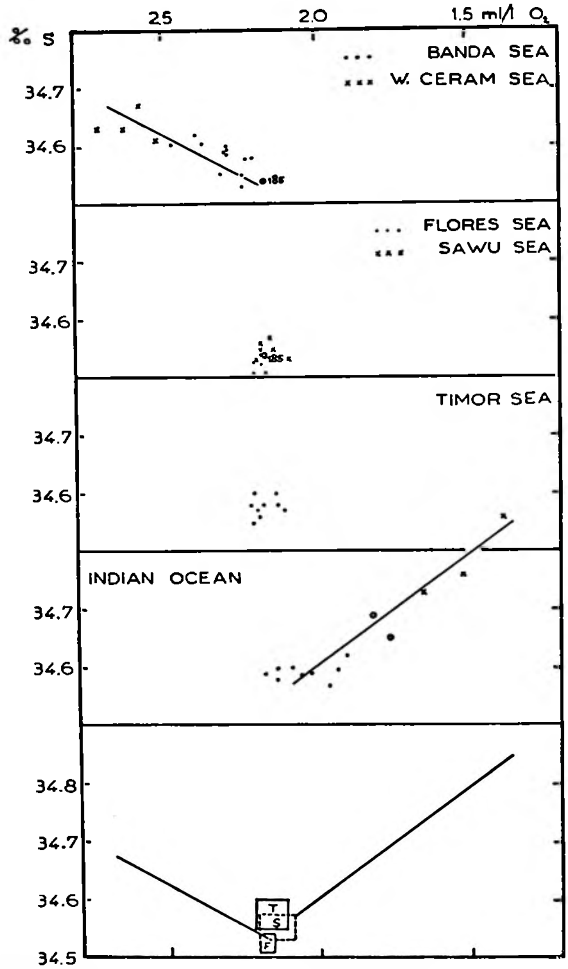


Fig. 8 A. Relation between salinity and oxygen content at a depth of 600 m, corresponding with the depth of the centre of the oxygen minimum in the southern part of the area investigated. Station 185 represents conditions on the threshold between the Flores Sea and Makassar Strait; the other dots and crosses are based on observations at stations of the vertical sections II and IV; in the Indian Ocean crosses represent observations at the "Dana"-stations 3804, 3805 and 3807 south of Java (THOMSEN, 1937); the circles represent observations of the "Albatross"-stations 183 and 190 (BRUNEAU, JERLOV and KOCZY, 1953) in the same area.

A centre with a comparatively low oxygen content and a comparatively high salinity, occurring at a depth of about 500 m in the northern part of the section, corresponds with the intermediate water mass south of Java, already described above. In this water the oxygen content is higher and the salinity lower when the component of Indonesian water becomes more important. According to this point of view, more Indonesian water is present in the southern part of the profile than in the northern part. This distribution is in accordance with the fact that the main passages communicating with the Indonesian seas are located in front of the southern part of the profile.

The course of the isopleths gives the impression, that down to a depth of about 900 m the distribution of the two factors considered is determined by the relative proportions of Indian Ocean and Indonesian water in the mixture. Below this depth the distribution is more uniform, while the oxygen content gradually increases from north to south; the highest values occur in station 131, located in front of the Timor Sea.

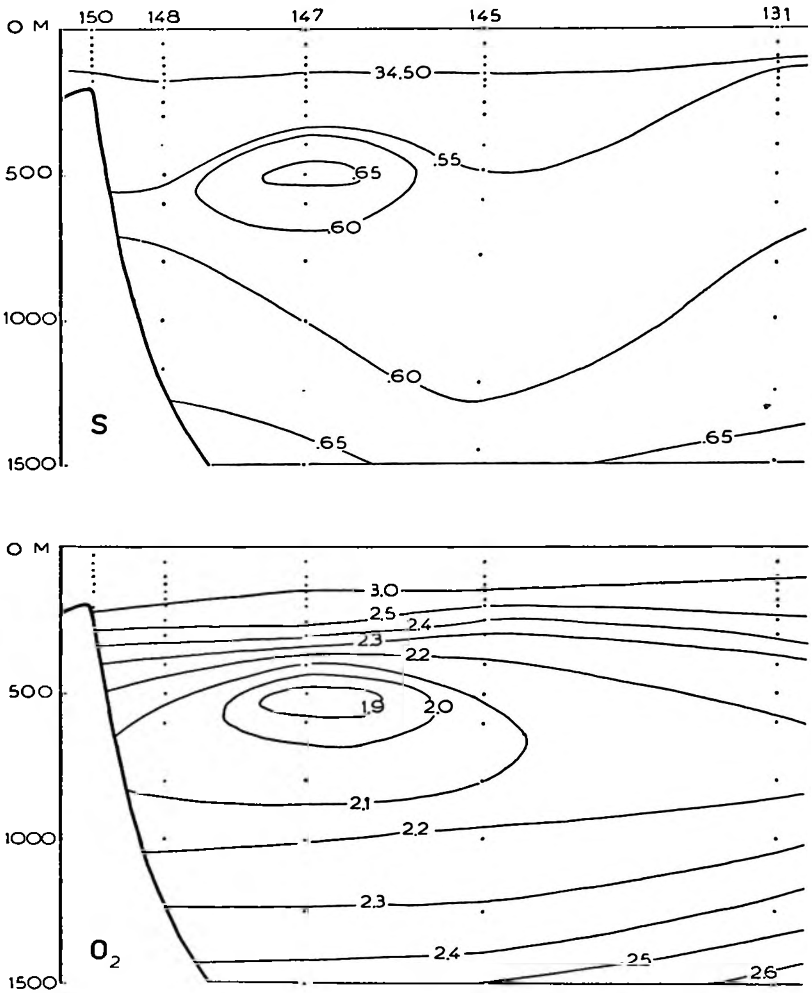


Fig. 8 B. Salinity and oxygen content in a north-south profile in the Indian Ocean; only the observations above 1500 m are represented.

From this station in the direction of the inner part of the Timor Sea the oxygen content in the water layer between 1000 and 1500 m gradually decreases (section II). Hence the increase towards the south along the profile considered cannot be due to the supply of Indonesian water, but must be caused by the gradual increase of oxygen content from south to north in the Indian Ocean itself.

This means, that in the layers below 900-1000 m a water mass occurs in which Indian Ocean water prevails. Hence, the decrease of oxygen content in the Timor Sea in a direction away from the Indian Ocean, although very slight, must be due to the penetration of Indian Ocean water. This penetration, however, is restricted to the Timor Sea itself, since the oxygen content in the Banda Sea is again slightly higher than in the Timor Sea (compare the horizontal distribution of oxygen at 1250 m, fig. 23H).

All things considered the influence in the eastern part of the Indonesian Archipelago of water from the Indian Ocean seems to be confined to its extreme southern part. However, it is necessary to

substantiate this conclusion in the future by further oceanographic investigations, especially in the region west of Australia. Such an investigation could at the same time establish in detail the distribution of Indonesian water, i.e., in the last instance, the dispersal of Pacific water in the Indian Ocean.

As regards the oxygen content in the water masses below the thresholds, VAN RIEL (1943) has already given attention to the oxygen content of the bottom water. A number of remarks in relation to this investigation have been made in chapter II, A and no further attention to this subject need be given here.

The same author observed that the oxygen content of the deep water in the basins is primarily determined by the oxygen content of the water flowing over the sills. For the oxygen content of a basin the depth of the intake is therefore very important. The remark has already been made on p. 36 that the Celebes Basin is supplied with water from the Pacific oxygen minimum at 1000 m. It is not surprising, therefore, that with the exception of the Sulu Sea the oxygen values in the deep water of the Celebes Sea are the lowest of all basins. The oxygen content near the intake amounts to about 2.2 ml/l. All other basins have a higher oxygen content since their saddle depths are either above the oxygen minimum (Halmahera Basin) or below it (Molucca Sea Basins, Buru Basin, complex of Banda Basins, Timor Basin, Aru Basin). These features need not be considered in detail, since they will become clear at once from an inspection of the vertical sections.

The Sulu Sea conditions are not very clear owing to a lack of observations near the sill of the China Sea. Observations of the "Dana" expedition (THOMSEN, 1937) show that in the latter sea at the saddle depth of 400 m the oxygen content amounts to 1.6 ml/l, a value which closely corresponds with those found at the same depth in section VI.

2. *Horizontal sections.*

50 m (fig. 23A, p. 104).

At a depth of 50 m the seasonal influence on the movements and properties of the water masses should be taken into account and the oxygen distribution must therefore be represented for each season separately. Only for the period from June-November are a sufficient number of data available. For this part of the year we find that in the Pacific Ocean and the greater part of the Molucca and Celebes Sea on the one hand and in the Indian Ocean and the Timor Sea on the other, the oxygen values are above 4.5 ml/l (at 27° C the saturation value amounts to 4.75 ml/l). In the greater part of the Archipelago itself, however, the oxygen content is somewhat below 4.5 ml/l and in the eastern part of the Banda Sea and in the Arafura Sea the oxygen values are even considerably lower, down to 3.0 ml/l. Other areas with a comparatively low oxygen content are located in the Flores Sea and the Sawu Sea and on the sills between the Sulu Sea and the Celebes Sea, between the Celebes Sea and the Molucca Sea, and between the Halmahera Sea and the Ceram Sea.

When the distribution of oxygen at 50 m is compared with the distribution of temperature in the same layer it appears that a comparatively low oxygen concentration is mostly accompanied by a comparatively low temperature. Therefore, the comparatively low oxygen values at 50 m depth of the inner parts of the Indonesian area are probably not due to an increased consumption of oxygen, but they may be caused by the fact that inside the Archipelago an increase of vertical mixing or, more generally, an increase of the influence of water from deeper layers causes at the depth under consideration a decrease of both temperature and oxygen content.

The Celebes Sea has a somewhat higher oxygen concentration than the Halmahera Sea. Evidently the South Equatorial Current has a lower oxygen content than the North Equatorial Current. This difference, which is even more pronounced at 100 m depth, provides an important means to trace the distribution of the water masses of both currents (compare p. 73).

100 m (fig. 23B).

Also at a depth of 100 m the distribution of oxygen will not be quite the same the whole year around. In this connection wo only have to point to the seasonal variations in the Pacific equatorial currents, which are felt down to a depth of about 150 m. If, nevertheless, all oxygen data are used in one chart, without regard to seasonal influence, this is done because the seasonal variations do not seem to influence the main characteristics of the oxygen distribution and because the number of data otherwise would be too small for a good representation.

In the greater part of the area the oxygen concentration at 100 m is below 4.0 ml/l; only in the Pacific Ocean near Mindanao and locally in the Celebes Sea is the concentration higher.

This is due to the fact already mentioned that the water masses near and in the maximum layer of the North Equatorial Current have a higher oxygen content than those of the South Equatorial Current (compare p. 37).

In the eastern Banda Sea and the Arafura Sea low oxygen contents are still more or less accompanied by low temperatures, but the main feature of oxygen distribution is the decrease in oxygen concentration with increasing distance from the Pacific Ocean.

The lowest oxygen contents of the whole area occur in the Sulu Sea, where water poor in oxygen enters from the China Sea. The course of the isopleths over the sill between the Sulu Sea and the Celebes Sea suggests that transport of water over this sill from one sea to the other is of little importance.

200 m (fig. 23C).

The oxygen distribution at 200 m is more complicated than at 100 m; this is in the first place due to the appearance of an area of low oxygen content (down to 2.5 ml/l) north of the Talaud Islands. The decrease in oxygen, which is accompanied by a decrease in temperature and salinity, is caused by the ascent of water poor in oxygen at the boundary of the North Equatorial (Mindanao) Current and the Equatorial Counter Current (p. 18).

The comparatively high oxygen content in the Celebes Sea, which is accompanied by a comparatively high salinity and temperature, reflects the conditions of the North Equatorial Current. The Current follows a course through the northern part of the Celebes Sea and enters this area close along Mindanao.

The water rich in oxygen flows towards the south into the Strait of Makassar, gradually decreasing in oxygen content. There is an indication that in the southern part of this strait and in the Flores Sea the water with the highest oxygen content—and the highest salinity and temperature—is present on the left hand side of the current, as is to be expected south of the Equator, where the Coriolis force has the tendency to push the warmer (lighter) water to the left.

The oxygen concentration of the water which enters the western Banda Sea from the Flores Sea, is still high (3.0 ml/l) as compared with the water flowing into the Banda Sea from the Ceram Sea (2.8 ml/l). The low oxygen content of the latter area and also that of the Molucca Sea and the Halmahera Sea is caused by the presence of much South Equatorial water.

A consequence of the difference in oxygen concentration between the water from the Ceram Sea and that from the Flores Sea is, that the oxygen distribution in the Banda Sea, just as the distribution of temperature and salinity, gives an insight into the transport of the two water masses in this area. The Ceram Sea water enters the Banda Sea north as well as west of the island of Buru. The same holds for the other layers above the threshold between Buru and Ceram, which has a saddle depth of about 1200 m.

Comparatively high oxygen concentrations are further found in the Sawu Sea and west of the island of Timor, where the Flores Sea water passes on its way to the Indian Ocean. On the whole, however, the oxygen content clearly decreases from the Pacific towards the Indian Ocean. This decrease is caused by oxygen consumption in the water during its flow towards the south (compare p. 41). The western part of the Timor Sea, and the Arafura Sea, lie beside the main route of water transport; the oxygen values in these areas are therefore exceptionally low, a situation which is also found in the other water layers.

400 m (fig. 23D).

The distribution of oxygen at 400 m differs considerably from that at 200 m: in the Pacific Ocean the highest oxygen values are no longer found near Mindanao, but near New Guinea. This is due to the fact that below the strata of the Equatorial Currents the oxygen content in this ocean decreases from south to north.

In contrast to what is found at 200 m, the Celebes Sea and the Strait of Makassar have therefore at 400 m a lower oxygen contentration than the Halmahera Sea and the Ceram Sea. As a result also the oxygen distribution in the Banda Sea has become reversed, the water with the highest oxygen

content now occurring in the northern part. The Makassar Strait-Flores Sea water flows through the Banda Sea to the east as far as the Arafura Sea.

Like the corresponding salinity distribution, also the oxygen distribution at 400 m suggests that through the Strait of Lifomatola there is no substantial flow of water to the south. The area of high oxygen content north of the island of Buru, however, is not directly connected with the corresponding area in the Halmahera Sea, since between these two areas the oxygen content is slightly lower. This separation was also observed in the case of the salinity distribution. The increase in salinity north of Buru was explained there as the result of vertical mixing (p. 19), which will at the same time cause an admixture of water with higher oxygen content from higher water layers.

The distribution of oxygen in the Celebes Sea suggests that water from this area flows into the Molucca Sea through the passages north of Celebes. Also the distributions at 100, 200, 600 and 800 m give some indications of such a transport, which would imply a counterclockwise water circulation in the Celebes Sea.

Near Mindanao the level under consideration intersects the nuclei of low oxygen content already discussed previously. Other areas poor in oxygen are the Sulu Sea and the Arafura Sea. In the Indian Ocean, finally, the oxygen content is also low, especially near the Lesser Sunda Islands. The causes for the presence of low oxygen values in these areas have already been examined above.

600 m (fig. 23E).

Only a few words need be said about the conditions at this depth, since the oxygen distribution is on the whole identical with that at 400 m.

The connection between the Halmahera Sea and the Ceram Sea is interrupted, so that all water entering the Banda Sea from the north has to pass through the Molucca Sea. Transport of water through this area, as we saw above, is slow. This is also reflected in the fact that in the southern Banda Sea and adjacent areas by far the greater part of the water originates from the Strait of Makassar (p. 31).

800 m (fig. 23F).

The connection between the Strait of Makassar and the Flores Sea is interrupted at 650 m and the only open connection of the Banda Sea with the Pacific Ocean at 800 m is formed by the Strait of Lifomatola. The oxygen content still decreases from this area towards the more remote parts of the Banda Basin system, especially towards the Flores Sea and the Bali Sea. It seems reasonable to ascribe this decrease to oxygen consumption in the water masses during their flow away from the entrance. However, the salinity data indicate, that water from the Strait of Makassar at 800 m still influences the conditions in the Flores Sea and the western and southern Banda Sea. This water mass has a comparatively low oxygen content (2.2 ml/l) and the low oxygen concentration in the areas considered may therefore to a large extent be due, not to oxygen consumption, but to a continued inflow of oxygen-poor water from this source.

1000 m (fig. 23G).

The oxygen distribution at 1000 m is not essentially different from that at 800 m. In most areas the oxygen contents have further decreased, since the second oxygen minimum of the Pacific Ocean has its centre at this depth (p. 29). In the seas adjacent to the Indian Ocean, however, a slight increase in oxygen content has taken place; the influence of Indian Ocean water is probably partly responsible for this increase.

The oxygen distribution in the Banda Sea suggests that, besides west of Buru, substantial amounts of water enter this area through the passage between Buru and Ceram; beyond this inlet the water turns to the left.

1250 m (fig. 23H).

The connection between the Indian Ocean and the Sawu Sea is interrupted at this depth and the remaining connections between the two oceans have become very narrrow. The western Timor Sea is filled with water from the Indian Ocean (p. 33). The direct flow of water from the Pacific Ocean to the Indian Ocean has therefore come to a standstill and water movements approach the conditions occurring below the thresholds of the basins.

The main entrance of new water for the system of Banda basins can be observed in the western Ceram Sea (Buru Sea), where the oxygen content is comparatively high. From this place the water is slowly transported to all parts of the Banda Sea and during this flow the oxygen content gradually decreases owing to oxygen consumption. The lowest values are found in the remote Bali Sea (2.1 ml/l). Also in the Sawu Sea the oxygen content is low. In the Arafura Sea, however, the oxygen content does not decrease very much, since water renewal takes place both from the side of the Timor Sea and the Ceram Sea. The passage between Buru and Ceram is closed.

2000 m (fig. 23I).

All important basins, with the exception of these in the Molucca Sea, are at this depth separated from the open ocean and their oxygen content is in the first place determined by the saddle depth of the deepest inlet.

For the Celebes Basin this inlet is located close to Mindanao (VAN RIEL, 1934). However, close to Celebes a second passage (Kawio Strait) exists, which has but a slightly shallower threshold depth. The oxygen distribution in the western Celebes Sea at 2000 m and also at 3000 m suggests that perhaps some deep water also flows over the latter sill.

With increasing distance from the main intake, the oxygen content in the basins gradually decreases. Attention must be drawn, finally, to the much greater oxygen content of the Indian Ocean as compared with the Pacific Ocean (3.2 against 2.7 ml/l). The cause of this difference is the greater distance of the Pacific deep water from the source in the Antarctic Ocean.

3000 m (fig. 23K).

The 3000 m level does not add new details to the picture of oxygen distribution presented above, but it is nevertheless reproduced here as a link between the water layers above it and the bottom water. The oxygen distribution of the bottom water itself is not dealt with in this contribution, since it has been described already by VAN RIEL (1943).

At the depth considered also the basins of the Molucca Sea are separated from the Pacific Ocean. Moreover, the Sawu, Wetar and Flores Basins are separated from the Banda Sea. In all these basins the oxygen content is mainly determined by the concentrations in the water masses entering over the thresholds.

3. *Temperature-oxygen relations.* The use of curves representing the relation between temperature and oxygen sometimes provides a valuable means to characterize water masses. Moreover, since oxygen is subject to changes by production and utilization in situ, comparison of temperature-oxygen curves can lead to a better insight into the magnitude of these changes. It is for both reasons that temperature-oxygen curves of the area studied have been drawn.

To facilitate a comparison with the TS-relations, practically the same groups of stations have been used to construct the TO_2-diagrams; table 9 gives a review of the areas and stations selected (compare table 8, p. 26). The temperature-oxygen diagrams are shown in fig. 24 (pp. 114-116). Finally, these diagrams have been combined into figs. 9A-9H to show their mutual positions. The latter series of figures will here be discussed.

Fig. 9A represents the temperature-oxygen curves for the two oceans. Curves I and II represent the relation of the Pacific Ocean near Mindanao and New Guinea, respectively. Curve XVI represents the relation for the Indian Ocean.

The division of the Pacific Ocean data into two groups was necessary for the same reason as in the case of the TS-relations for this area (fig. 7A); curve I represents the conditions in and below the North Equatorial Current (Western North Pacific water) and curve II those in and below the South Equatorial Current (Equatorial Pacific water). The temperature-oxygen curves intersect at a temperature of 14° C (200 m); the North Equatorial Current has a higher oxygen content than the South Equatorial Current, while in deeper layers the oxygen concentration in general decreases from south to north. Curve II shows two oxygen minima; the upper minimum (22° C, 100-150 m) corresponds with the salinity maximum of the TS-curve (fig. 7A), the lower minimum (4.5° C, 1000 m) represents the oxygen minimum of the equatorial water. The position of the minimum of Curve I at 6.5° C is the result of the influence on the course of this curve of both the Equatorial oxygen minimum and the North Pacific oxygen minimum at 350 m (9° C).

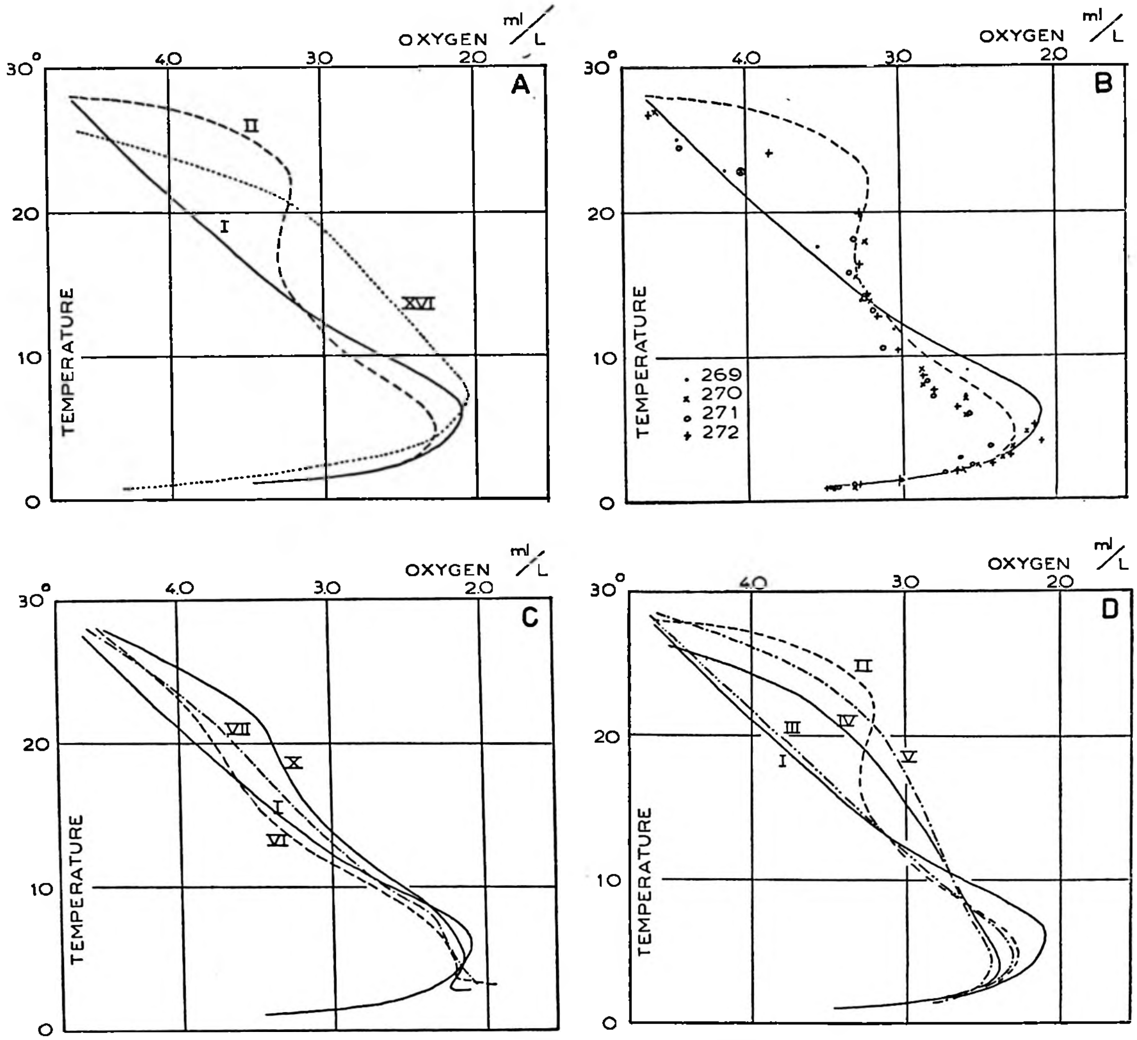
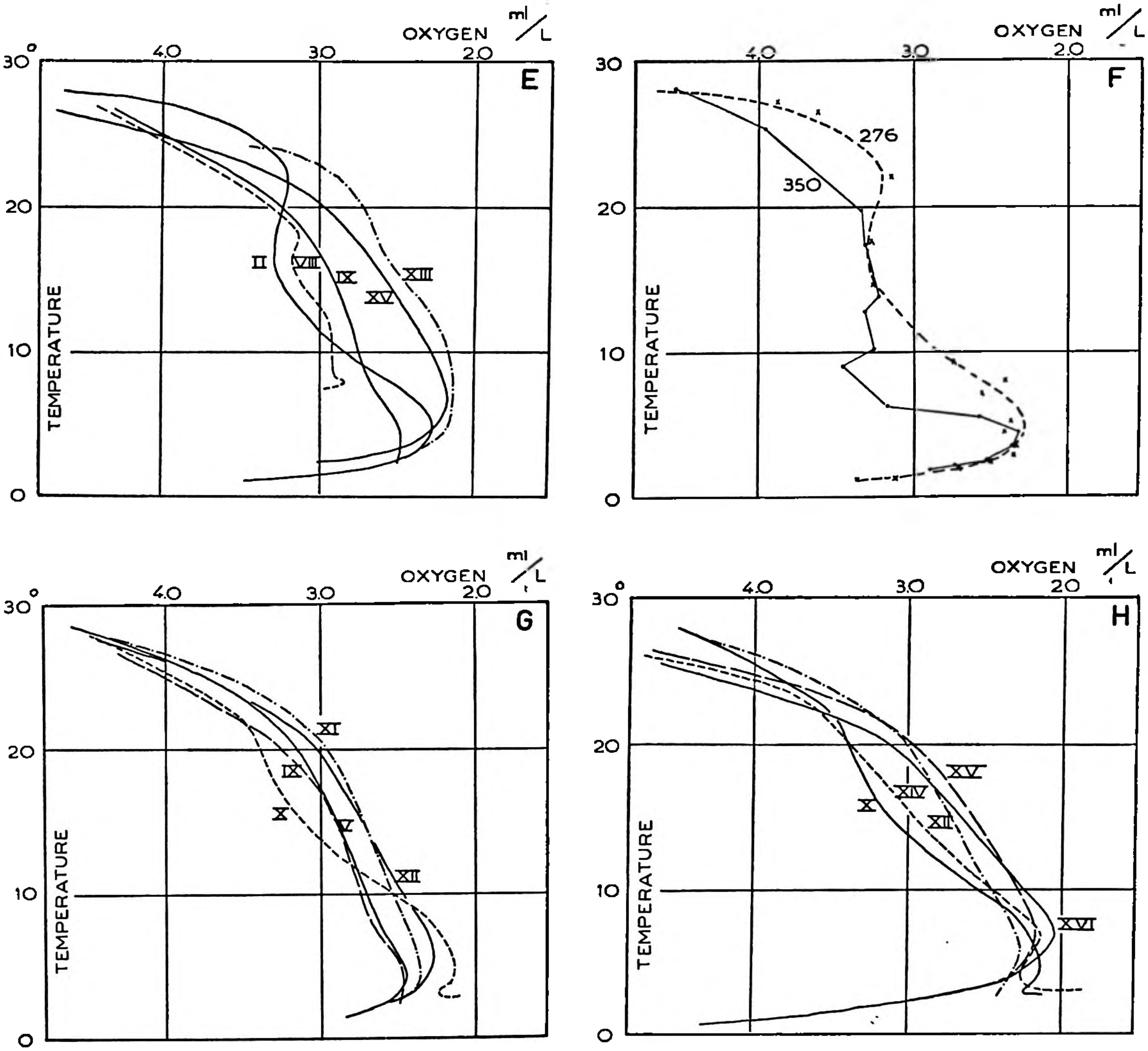


Fig. 9. Temperature-oxygen relations; compare table 9.

Curves with an intermediate character are found at the stations 269, 270, 271 and 272 (fig. 9B), where also the TS-relations show intermediate forms (fig. 9B). The boundary between the northern and southern water lies east of the Talaud islands; however, as has been observed before, the condition describe donly holds for the period between May and November, whereas in the other half of the year the boundary will be located farther to the SE.

The temperature-oxygen diagrams of the water masses flowing from the Pacific Ocean through the Celebes Sea and the Makassar Strait into the southern part of the Archipelago are shown in fig. 9C. The temperature-oxygen curve VI, representing conditions in the Celebes Sea, follows approximately curve I, representing the northern water. Towards the Flores Sea (curve X) there is a slight decrease in oxygen content; this decrease is most pronounced in the upper layers and in the deep water, but never exceeds the value of 0.4 ml/l (20° C, 100-150 m).



The curves for the water masses in the Molucca Sea are shown in fig. 9D. The temperature-oxygen relation III of the northern Molucca Sea follows in its upper part the northern Pacific curve I, thus indicating that in the upper layers northern water enters this sea. The lower part of curve III, however, corresponds with the southern Pacific curve II, since the intermediate and deep layers of the Molucca Sea are filled with Equatorial Pacific Water. In the upper layers there is an considerable decrease in oxygen content from the northern to the southern Molucca Sea (IV) and the Strait of Lifomatola (V); this decrease is greatest at 20° C (100-150 m depth), where it ammounts to 0,7 ml/l, showing that water transport is much slower in this area than in the Celebes Sea. In the deeper layers, below 9° C, the oxygen content increases from north to south.

The water masses near New Guinea are represented in fig. 9E, together with the Pacific temperature-oxygen relation (II). The Halmahera sea curve (VIII), which is based on the observations at only one station (nr. 353), occupied by the "Snellius" expedition in October 1930, differs considerably from curve II, based on observations at stations occupied in May. Seasonal fluctuations are

Table 9

Temperature-oxygen diagrams; compare figs. 9 and 24

Fig. 24	Area		Stations
A	I	Pacific Ocean, North Equatorial	262-266, 268, 293-296
B	II	Pacific Ocean, South Equatorial	273-276
C	III	Molucca Sea, North	283-288, 290-292
D	IV	Molucca Sea, South	80, 332-336
E	V	Lifomatola Strait	223-228
F	VI	Celebes Sea	47, 48, 52, 53, 56, 57, 75, 76, 301, 303, 304
G	VII	Makassar Strait	35, 39, 41, 310-312
H	VIII	Halmahera Sea	353
J	IX	Ceram Sea, West	229, 330
K	X	Flores Sea	167, 179, 180, 197
L	XI	Banda Sea, North	209, 210, 212, 215, 218, 331
M	XII	Banda Sea, South	205, 235, 238, 240, 241, 245, 246, 249, 319
N	XIII	Arafura Sea	100, 104
O	XIV	Sawu Sea	155, 159, 160, 163, 378, 379, 380
P	XV	Timor Sea	115, 118, 121, 125, 127
Q	XVI	Indian Ocean	145-147

responsible for this difference; this is also illustrated by the temperature-oxygen relations at stations 276 and 350, which lie in the same area and were occupied in May and October respectively (fig. 9F). The nature of these fluctuations has already been discussed in connection with the TS-relations. The comparatively high oxygen content in the intermediate water at the stations 350 and 353 corresponds with a comparatively high salinity of their TS-diagrams and must be due to the dominating influence of water from the southern Pacific. In the upper layers the comparatively high oxygen content, accompanied by a comparatively low salinity, points to influence from northern water near New Guinea already in October.

Curve IX for the western Ceram Sea (Buru Sea), like the corresponding TS-relation, is an intermediate form of the Halmahera Sea curve (VIII) and the diagram for the Strait of Lifomatola (V) (compare figs. 9D and 9E).

In the Arafura Sea (XIII, fig. 9E), the oxygen content is considerably lower; this decrease may to some extent be due to admixture of water from the west, but it must be mainly caused by oxygen consumption in the area itself: in the southern Banda Sea (XII, fig. 9G) and in the Timor Sea (XV) the oxygen concentration is again higher. The magnitude of the local oxygen loss amounts to 0.1-0.2 ml/l and decreases with depth.

The conditions in the Banda Sea (curves XI and XII), together with those of adjacent areas are represented in fig. 9G. If we first consider the latter areas, it appears that above 12° C (250-300 m) the Flores Sea (X) has a higher oxygen content than the Lifomatola Strait (V) and the western Ceram Sea (IX); below this depth the reverse is the case. This is due to differences in the oxygen content of the Pacific water masses supplied to these areas, which have already been described above.

The two curves for the Banda Sea intersect at about 200-250 m (14° C); above this depth the southern Banda Sea has a higher oxygen content than the northern Banda Sea; this must be due to an increasing influence from north to south of Flores Sea water, in relation to Ceram Sea water. At the same time, however, the oxygen content of the Banda Sea is about 0.2 ml/l lower than that of the areas of supply in both its northern and southern part. This decrease must be due to consumption of oxygen.

Below 9° C (400 m) the two Banda Sea curves lie between those for the Flores Sea and the western Ceram Sea. It has previously been shown that at 600 m the oxygen content of the Banda Sea water can simply be explained as the result of the intermixture of water masses from these two areas with-

out oxygen consumption being taken into account. The same probably holds for all water layers between about 400 m and 600-700 m: the sill depth between the Flores Sea and Makassar Strait.

Below the latter depth (6° C) all water in the Banda Sea and also most of the Flores Sea water originates from the Ceram Sea. The oxygen concentration gradually decreases from the latter area towards the Flores Sea; the maximum decrease amounts to about 0.3 ml/l. Below 3° C the Flores Sea is separated from the Banda Sea by a threshold and a further decrease in oxygen of 0.1-0.2 ml/l takes place in the isolated Flores Basin.

The temperature-oxygen curves of the regions south of the Banda Sea and the Flores Sea are, together with the curves for these two areas, represented in fig. 9H. The intermediate position of the Sawu Sea curve (XIV) between the curves for Flores Sea (X) and for the southern Banda Sea (XII) stresses the character of the Sawu Sea as a main passage for water from the North. Below 10° C (350 m) the loss of oxygen causes the Sawu Sea curve to be to the right of the other two curves; below 6° C (750 m) the isolated position of the Flores Sea causes its oxygen content to fall below that of the Sawu Sea; in the deep water, where the Sawu Sea also becomes isolated, the oxygen content of the latter area again decreases below that of the Flores Sea. The total decrease of oxygen content in the deep water of the Sawu Sea amounts to about 0.4 ml/l.

In the Timor Sea (curve XV) the oxygen content is at most temperatures higher than in the Sawu Sea, since the oxygen-decreasing influence of the Flores Sea water is of less importance in the former area. Below 7° C the oxygen content of the Sawu Sea becomes the lower of the two, not only because of the more isolated position of the Sawu Sea, but also since in the deeper levels of the Timor Sea Indian Ocean water penetrates with a comparatively high oxygen content.

The temperature-oxygen relation for the Indian Ocean, finally, is represented by curve XVI; this curve is also reproduced in fig. 9A, to facilitate a review of the total decrease of oxygen content in the water masses during their flow from the Pacific towards the Indian Ocean. In order to understand the course of curve XVI, it must first be realised that water passing through the Strait of Makassar plays a much more important rôle in the southern Indonesian seas than water passing through the other main inlets, at least above a depth of 600-700 m (6° C). Therefore the part of curve XVI above 6° must in the first place be compared with curve I of the North Pacific water. In the upper end the two curves approximately coincide and the oxygen content approaches saturation, since consumption of oxygen is neutralized by production due to phytoplankton and the uptake of oxygen from the air. Below 20° C (150 m) these influences are of no importance and the difference between the two curves will approximately represent oxygen consumption. At 150 m the difference in oxygen content between the two curves appears to amount to 0.7-0.8 ml/l. This difference gradually decreases with depth and is practically nil at 6° (600-700 m).

Below 6° C and above 4.5° C (1000 m) southern Pacific water will play the most important rôle. This is reflected in the fact that curve XVI crosses curve I at 6° C, but remains below curve II down to about 4.5° C. The oxygen decrease in the layers between 700 and 1000 m is difficult to determine, but certainly very small.

At 4.5° C (or 1000 m) curve XVI also intersects curve II and from this depth on shows the highest oxygen content of the three. Below 1000 m curve XVI practically represents Indian Ocean water without Pacific influences. The higher oxygen content of the Indian Ocean deep water if compared with the Pacific deep water may be assumed to be a consequence of the shorter distance of the former water mass from their common source in the Antarctic regions.

4. *Oxygen consumption.* The evident decrease in oxygen concentration of the water flowing through the East Indonesian Archipelago seems to offer an opportunity for the determination of oxygen content in a given water mass within a certain time interval. However, if one wishes to determine the oxygen consumption in unit time in various water layers, one should know the length of time in which the water passes from one ocean to the other. This period cannot be estimated correctly from the data available. Inspection of the few current measurements of the "Snellius" (LEK, 1938) shows, however, that the water at 150 m probably passes the Indonesian seas in less than one year, whereas at 600 m 5 years or more are needed. In deeper layers this period will be much longer still and, finally, in the deep water of the basins periods of a few hundred years may be expected.

The rapid decrease of current velocity with depth, combined with the decrease with depth of the difference in oxygen between the water masses discussed above, indicates that by far the greatest oxygen consumption takes place in the upper few hundred meters of the water column. In fact, at 150 m a difference in oxygen of 0.7-0.8 ml/l and a travel time of one year correspond with an oxygen consumption of 0.7-0.8 ml/l per year; at 600 m a difference in oxygen of 0.05 ml/l or less and a travel time of 5 years correspond with an oxygen consumption of only 0.01 ml/l per year or less. These data, although very approximative, are roughly in accordance with consumption values calculated by Riley (1951) for the Atlantic Ocean.

The conditions in the deep water of the basins need some further consideration. In one case the yearly oxygen consumption can be calculated approximately. For the complex of basins supplied through the Lifomatola Strait, we found in earlier pages that the rate of water renewal of the "homothermal" deep water can probably be estimated at about 300 years (p. 23). The oxygen content of the water entering over the sill of the Strait of Lifomatola amounts to 2.5-2.6 ml/l. Towards the extreme ends of the complex this concentration gradually decreases to 2.0-2.1 ml/l (Sawu and Flores Basins). The average oxygen content of the "homothermal" mass, however, amounts to 2.3-2.4 ml/l. If we take an average decrease in oxygen content of 0.3 ml/l to be the correct figure, the yearly consumption of oxygen in the deep water will amount to 0.001 ml/l. As the average depth of the "homothermal" water mass amounts to 2 km, this value corresponds with an oxygen consumption of 2000 ml/m²/year. An oxygen consumption of 1 ml/l is equivalent to the oxization of 0.536 mg carbon; hence the above value corresponds with the oxization of about 1 g carbon/m²/year.

The average gross production of organic carbon over all oceans was estimated by Steemann Nielsen (1954) at 0.15 g C/m²/day or 55 g/m²/year. This value also holds for the equatorial region of the Pacific Ocean; in the Banda Sea Steemann Nielsen and Jensen (1957) found a somewhat higher value (0.30-0.34 g C/m²/day), but this amount, according to the authors, may not be wholly representative for the area under consideration. Nevertheless the production in the Banda Sea may be somewhat higher than in the ocean. If we assume that the production lies between 55 and 110 g C/m³/year, 1-2% of this amount is decomposed in the deep water. This percentage, of course, is only a very rough approximation, but the order of magnitude may be right, and the method of calculation has the advantage of being a direct one, which cannot often be applied.

Data concerning the decomposition of organic carbon in the deep water of the oceans are scarce; Riley (1951, pp. 94 and 95) estimates the oxygen consumption in the deep water of the western North Atlantic Ocean below 1500 m at 0,0006-0,002 ml/l/year; this estimate was based on physical oceanographic calculations as well as on biological considerations. The value of 0,001 ml/l/year calculated for the Banda Sea is in good agreement with this amount.

An important question in connection with the consumption of oxygen in the basins is whether the rate of consumption is influenced by the depth of the threshold, i.e. by the water level from which the water in the basin originates. There seems to be more than one possibility: (1) the rate of oxygen consumption is mainly determined by the amount of organic matter raining down into the basins from the upper water layers; (2), oxygen consumption mainly depends on the amount of decomposable organic matter already present in the inflowing water.

In case (1) the rate of oxygen utilization will be mainly independent of the depth of the threshold. In case (2) there are two possibilities. The first, (2a), is that oxygen consumption in the deep and intermediate water is mainly caused by organic matter with a comparatively short time of decay. This material will show a distinct decrease in concentration with increasing distance from the water surface. The second possibility (2b) is that oxygen consumption in the deeper layers is mainly due to organic matter with a very long time of decomposition; this material, which to a very large extent consists of dissolved organic matter, is distributed evenly through the whole water column.

In case (2a), oxygen consumption in a basin with a high threshold must be significantly higher than in a basin with a low threshold. The Sulu Basin, for example, receives water from a depth of only about 400 m and its waters should have much greater oxygen requirements than the waters of other basins, for example the Banda Sea. However, the apparent oxygen utilization in the Sulu Basin is only 0.1 ml/l, from 1.6 ml/l at sill depth to about 1.5 ml/l in the "homothermal" deep (section VI, fig. 18F). The rate of water renewal in this basin will not be more rapid than that in the Banda Basin, since the Sulu Basin is surrounded by much higher thresholds than the Banda Basin.

Hence, also oxygen consumption in the Sulu Basin is certainly not faster than in the Banda Basin. This fact therefore, seems to exclude assumption (2a).

A decision between the possibilities (1) and (2b) seems not well possible on the basis of the data at present available. It must be observed, however, that in case (2b) the rate of oxygen consumption will be the same throughout the whole basin, whereas in case (1) it will decrease with depth (apart from consumption of oxygen by organic matter of the bottom sediment). Hence, if the differences in "age" between the various water masses inside a basin were exactly known, a decision between (1) and (2b) might be possible on the basis of the oxygen distribution observed.

The oxygen distribution in the various basins shows some interesting details. Whereas in the majority of the basins the oxygen content in the deep water increases towards the bottom, there are a few basins where the reverse occurs. The latter include the Sawu Basin, the Makassar Trough and, to a certain extent, the Flores Basin.

If we could assume that the oxygen concentration inside a basin is mainly a function of the "age" of the water, an increase in oxygen towards the bottom would indicate that the "newest" water in the basin is present close to the bottom. In fact, much of the water entering a basin over the threshold will flow down towards the bottom along the inner slope of the intake and from there gradually move upward (p. 55).

The decrease in oxygen towards the bottom in the three above-mentioned basins might indicate that the water near the bottom is older than the water from the higher levels; it might also mean that the bottom water has been in contact with the bottom for a long period, so that much oxygen has been utilized by the bottom sediment. Which of these two assumptions may be correct cannot be decided at present; however, both point to a comparatively high "age" of the water in the deeper parts of the basins in question. We shall return to this problem in the next chapter in connection with the distribution of specific alkalinity in the basins.

B. *Hydrogen ion concentration and specific alkalinity.* The hydrogen ion concentration and the specific alkalinity are discussed together in the same section, since both are of importance for the same problem: the solution and sedimentation of lime. This problem has been briefly outlined in the introduction.

The investigations of the "Siboga" and "Snellius" expeditions brought to light exceptional sedimentary conditions in the East-Indonesian Basins. The average lime content appeared to be considerably lower than in the open oceans.

The actual position of the basins can at present be determined more accurately than in the years of the expedition, since the lime contents of especially the sediments in the Pacific Ocean are now much better known. In fig. 10 the average relation between calcium carbonate content and depth of the sediments on the floor of the Indonesian basins, according to KUENEN (1943) and NEEB (1943), is compared with the relation at different latitudes in the Pacific Ocean, according to REVELLE (1936). We see that in this ocean the lime content of the ocean floor decreases towards the north and that the conditions of the basins correspond most with those in the Pacific Ocean between 10 and 20° north. Compared with the Pacific Ocean, the lime content in the sediments of the basins is therefore not exceptional; only if this content is compared with the part of the Pacific Ocean in the same latitude, 5° South to 5° North, is there a considerable difference.

Several suggestions have been made to explain this difference, such as a better solubility of lime in the water of the basins, a fast rate of renewal of this water, and a possible "dilution" of the sediments with terrestrial material from the surrounding islands.

These possibilities will be discussed in more detail below. First, however, the main characteristics of the pH- and alkalinity distribution should be described. The knowledge of the hydrogen ion concentration will make it possible for us to estimate the lime dissolving capacity of the water masses, while the alkalinity data will enable us to determine whether lime is actually dissolved or not.

1. *Hydrogen ion concentration.* For a representation of the distribution of pH a number of vertical sections have been drawn, which correspond with those constructed for the chemical properties already discussed (figs. 19 A-F, pp. 80-84). The inaccuracies in the pH-determinations, mentioned in chapter II B, essentially influence the course of the isopleths in a number of cases. Nevertheless, no

corrections have been made for these errors, since only by reproducing the pH-distribution on the basis of the values determined can the influence of the discrepances be correctly interpreted. Further, the pH-observations are represented without subtraction of the decrease caused by pressure.

Changes in the hydrogen ion concentration of a given water mass are due to several factors, of which the change in the amount of carbon dioxyde is the most important. Production of carbon dioxyde is closely connected with the consumption of oxygen and in the water layers not in immediate contact with the air a relation between the distribution of oxygen and the pH is therefore to be expected, a low oxygen content corresponding with a low pH.

A comparison of the distribution of these properties in the vertical sections in general confirms this expectation. In section I (fig. 19 A) the pH-nuclei at a depth of 400 m correspond with the isolated areas of low oxygen content found at the same depth (fig. 18 A). Also the low oxygen content of the water flowing over the sill of the Celebes Basin is reflected in the comparatively low pH of this water. In the deep water of this basin the pH decreases in the direction of the Makassar Strait and the same holds for oxygen.

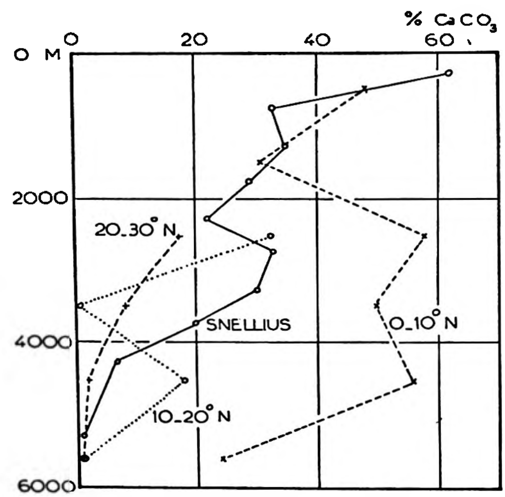


Fig. 10. Average relation between depth and lime content of the sediments in the Indonesian basins; according to Kuenen (1943) and Neeb (1943). Average relations at different latitudes in the Pacific Ocean according to Revelle (1936).

In section II (fig. 19 B) the pH-minimum in the Pacific Ocean coincides with the oxygen-minimum there. In the Buru- and Banda Basins the isopleths follow a capricious course, due to the extremely small differences of pH in the deep water, which lie closely above or below 7.8. In the Timor Sea, the pH-minimum again corresponds with the oxygen-minimum. The values in the Timor Sea may be somewhat too high if compared with those more to the north, since, as has been shown in chapter II B, there is a small discrepancy between the pH-estimations during the first and second part of the expedition.

The minima in section III (fig. 19 C) correspond with the oxygen-minima in this section. Moreover, the comparatively high oxygen content of the water entering the Halmahera Sea is reflected in the comparatively high hydrogen ion concentration in this area. The pH-measurements on the right hand part of section III may be somewhat too high if compared with the left hand part.

In section IV (fig. 19 D) the isopleths in the deep water follow the same capricious course as in section II. It is somewhat surprising to find that in the deep water of the Flores Basin, contrary to expectation, the pH increases with decreasing oxygen content. We shall see afterwards that in this basin the specific alkalinity is also exceedingly high, so that the exceptional rise of pH may be explained from an increase of specific alkalinity. The pH-values in the left hand part of section IV

are probably, for the same reason as in section II, somewhat too high in comparison with those in the right hand part.

In the Sawu Basin the discrepancies between the left and the right hand parts of section V are even greater than in other sections, due to errors in the pH-determination, so that the stations 146-162 of section V cannot be used in relation with 373-377. For this reason the distribution of pH in section V has not been reproduced in this paper. The values at the stations 373-377 (table 1, Vol. IV), which show a pH in the deep water of the Sawu Basin somewhat below 7.80, are probably the most accurate ones (see Chapter II, B); the data of the stations 146-162 must probably be reduced by 0.05. After this correction the pH in the deep water of the Sawu Basin locally shows a pH of 7.75. This comparatively low value corresponds with the low oxygen content of the deep water in the Sawu Basin. At intermediate depths, low pH-values on the side of the Indian Ocean (stations 146-162) correspond with low oxygen concentrations.

Section VI (fig. 19 E) follows a course through the Sawu Sea and the Celebes Sea from north to south and intersects section I in the latter area. Comparison of the pH-values in both sections immediately shows those in section VI to be too high compared with section I. Inspection of the original data indicates that the difference amounts to about 0.07 pH-units. Hence, the pH estimated in the deep water of the Sulu Basin will also be too high, the actual pH probably being close to 7.80, instead of being 7.85-7.90, as shown in the diagram.

The pH of the deep and bottom water in the area investigated is low as compared with the corresponding water masses in the oceans. In the Atlantic Ocean the pH, without pressure correction, is nearly always above 8.0 (Wattenberg, 1933), whereas in the Indonesian basins a value of about 7.8 is encountered. In the part of the Indian Ocean near the area investigated the pH is continually above 7.9, while in the Pacific Ocean pH-values are always above 7.85.

The conditions in the latter ocean need our special attention in view of its predominant influence on the waters of the Indonesian seas. The observations of the "Carnegie" and "Albatross" expeditions show that in the northern Pacific Ocean minimum pH-values down to 7.5 occur in the oxygen minimum layer, whereas in the deep and bottom water the pH is between 7.8 and 7.9. Towards the south the pH in the deep water of all oceans approaches the value of 7.9-8.0, occurring in the water masses around the Antarctic continent.

The low pH of the water in the Indonesian basins is partly due to the fact that the water masses in most basins originate from the Pacific Ocean, partly to the circumstance that these water masses are derived from layers in or not far from the centre of the intermediate pH minimum.

Table 10 summarizes the pH-values *in situ* at different depths in the deep water of the various basins and both adjacent oceans.

Table 10

Hydrogen ion concentration *in situ* [1])

	2000 m	4000 m	6000 m	Maximum depth	
Pacific Ocean	7.75	7.75-7.80	7.70-7.75	7.66	(10500 m)
Indian Ocean	7.75	7.75-7.80	7.75-7.80 [3])	—	
Sulu Basin	7.75	7.75	—	7.72	(5580)
Celebes Basin	7.75	7.70	7.65	7.65	(6220)
Makassar Trough	7.75	—	—	7.73	(2540)
Sangihe Trough	7.75-7.80	—	—	7.70-7.75	(3820)
Halmahera Basin	7.80	—	—	7.80	(2039)
Moluccan Sea Basins [2])	7.75-7.80	7.70-7.75	—	7.70	(4810)
Buru Basin	7.70-7.75	7.70	—	7.65	(5319)
Banda Basins	7.75	7.70	7.65-7.70	7.60	(7440)
Flores Basin	7.75	7.70-7.75	—	7.70	(5130)
Sawu Basin	7.75	—	—	7.65	(3470)
Aru Basin	7.85	—	—	7.70	(3680)
Timor Trench	7.75-7.80	—	—	7.75	(3310)

[1]) corrected for experimental errors [2]) Batjan Basin [3]) Deepest "Snellius" observation

2. *Specific Alkalinity*. a. *Reliability of the observations.* Also in the case of specific alkalinity vertical sections have been constructed, mainly corresponding with those drawn for the other chemical properties (figs. 20 A-C, pp. 85-87). Before considering these sections it is necessary to discuss the reliability of the alkalinity data. Unlike the data for oxygen and pH, no duplicate stations from which methodical errors could be directly estimated, are available for specific alkalinity. In this case, to test the reliability of the results, all data have to be compared with each other. For this reason the following discussion finds a better place in this chapter than in chapter II.

The data of all stations located in one basin have been used to construct a specific alkalinity-depth diagram for this particular basin. The stations used are summarized in table 11; the various diagrams are represented in figs. 11 A-M.

Table 11

Specific Alkalinity-Depth diagrams; compare figs. 11 and 12.

Fig. 11	Basin	Stations
A	Pacific Ocean	262, 264, 265, 271, 272, 291, 296, 350
B	Sulu Basin	64, 66
C	Molucca Sea Basins	337, 340, 342, 344, 345
D	Celebes Basin-Makassar Trough	33, 41, 46, 52, 74, 76, 77, 301, 303, 310, 311
E	Halmahera Basin	353
F	Buru Basin	90, 229, 328, 330, 354A, 355
G	Flores Basin	149, 166, 167, 168, 173, 175, 178, 179, 180, 187, 189, 191, 192, 193, 194, 197, 198, 318
H	Banda Basins	202, 203, 205, 208, 209, 210, 215, 216, 235, 241, 246, 319, 321, 356, 258, 359, 362, 365, 369, 373, 374, 376
J	Aru Basin	94, 95, 99, 104, 105
K	Sawu Basin	153, 154, 155, 158, 159, 163, 379, 380
L	Timor Trench	109, 112, 115, 118, 120, 121, 122, 127, 128
M	Indian Ocean	131, 141, 144, 145, 146, 147, 148

Starting with the Celebes Sea-Makassar Trough group, we see in fig. 11 D that the series of stations 41, 46, 52, 74, 76 and 77 (dots) show at every depth considerably higher alkalinity values than the series 301, 303, 310 and 311 (crosses). This difference is nearly constant from the surface down to the bottom; one series contains stations of the first part of the expedition only, and the other series only those of the latter part. These facts can only be explained by assuming a systematical error of the alkalinity determination during the first or second part of the expedition. An exception must be made for station 33 (crossed circles), the values of which lie between the results of the two series.

In the Buru Basin only one station (nr. 90; crosses in fig. 11 F) was occupied in the early part of the expedition, while the other stations are from the middle and latter part of the investigations. The values of station 90 are higher than those of the stations occupied afterwards, but the difference between the observations is not so large as in the case of the Celebes Sea.

In the Aru Basin (fig. 11 G), only alkalinity data from the first part of the expedition are available (nrs. 94-105). The specific alkalinity of the surface values of this series on an average amounts to 0.1250; this is a conspicuously higher figure than that found in other areas, where specific alkalinities seldom exceed 0.1225. Hence, we must assume that also the values of the Aru Basin series are too high in comparison with those of the latter part of the expedition.

Also in the Timor Trough and Indian Ocean (stations 109-148, figs. 11 L and 11 M) samples were taken only during the first part of the expedition, so that no direct comparison with data from the latter part is possible. In the Flores Sea, however, the alkalinity values of station 149 (fig. 11 I, crosses) are not essentially different from those of the near-by station 318 (circles) and also in the Sawu

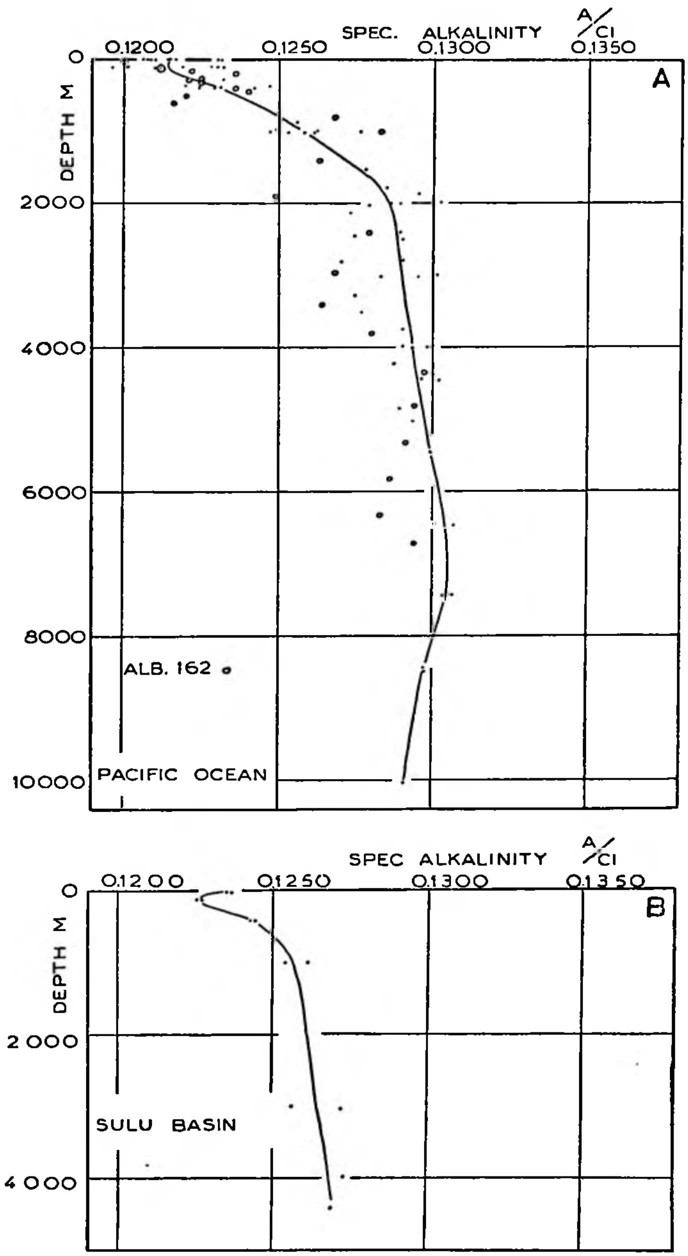


Fig. 11. Vertical distribution of specific alkalinity; un corrected data (see text).

Sea (fig. 11 K) the near-by stations 155 and 379 show about the same alkalinity values. The data of the remaining basins (figs. 11 A, C, E and H), except the Sulu Basin (fig. 11 B), likewise show no methodical errors. Therefore, at least from station 149 on, all data are mutually comparable. Probably, however, already the Timor Sea data belong to the latter group, since the surface values in this area

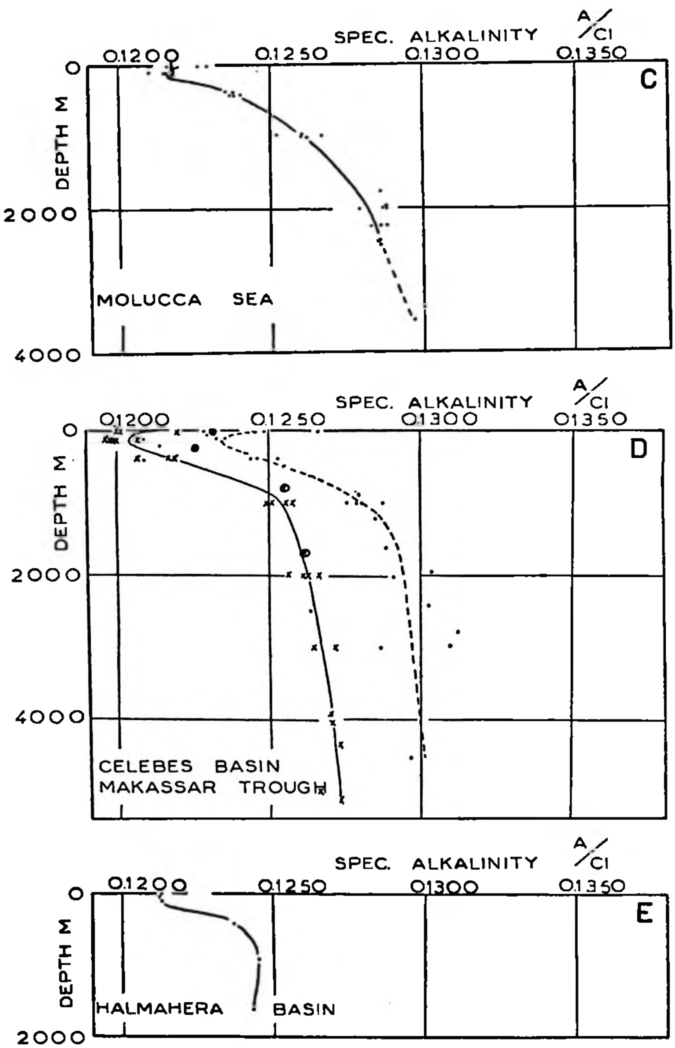


Fig. 11· C-E

and in the Indian Ocean are much lower than those in the Aru Basin and approximately the same as those in the Sawu Sea; moreover, they are only slightly higher than those in the Banda and Flores Seas and for this small difference a sound oceanographical reason can be advanced (p. 57).

As regards the absolute accuracy of the alkalinity data, the "Snellius"-measurements may be compared with those of the "Albatross" expedition (Bruneau, Jerlov and Koczy, 1953). Two stations of the latter expedition were occupied in the area of investigation, one in the Pacific Ocean (nr. 162) and one in the Flores Sea (nr. 173). It appears from figs. 11 A and I, that the "Snellius"-measurements are slightly higher than the Swedish observations; however, on the whole the agreement is not unsatisfactory, especially if one realizes that the two expeditions followed quite different procedures for the determination of alkalinity. For the slight difference between the observations of the two expeditions no correction has been made and in the following pages the values of the second part of the "Snellius"-expedition are considered to be correct.

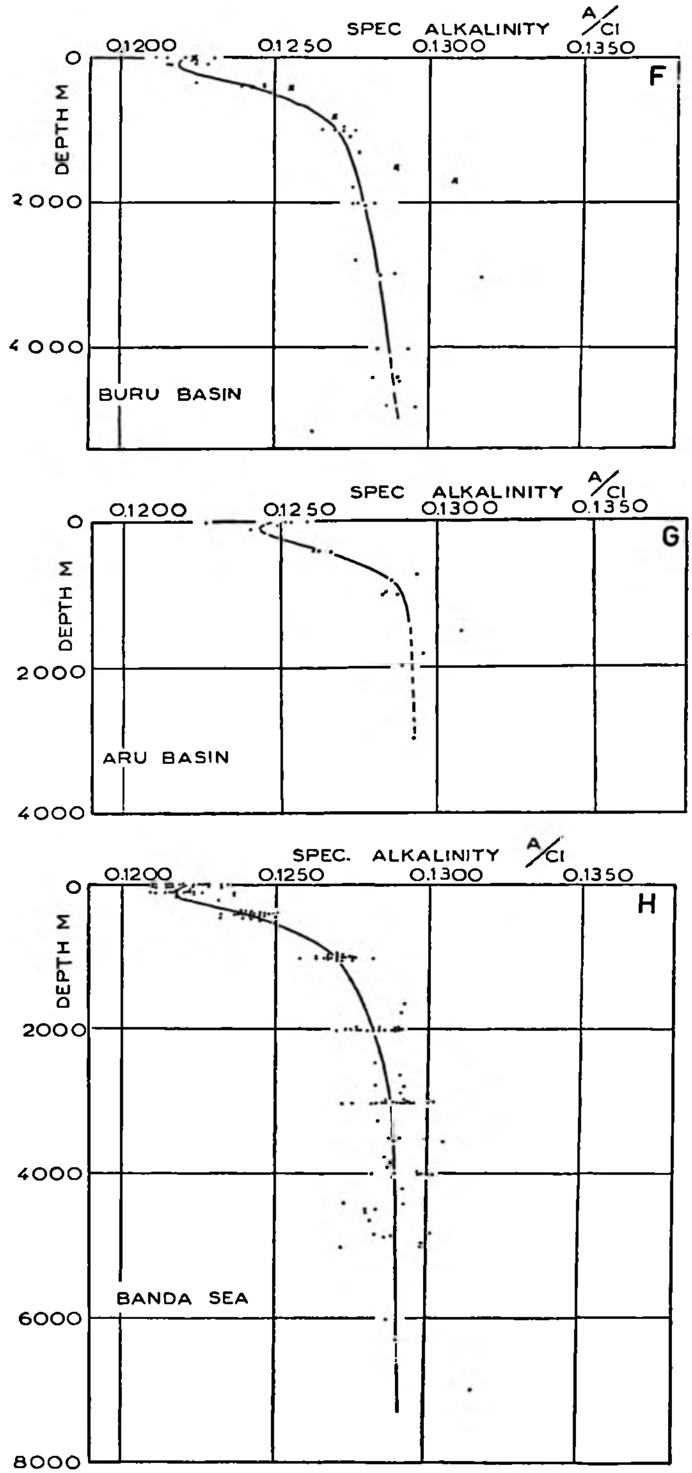

SPEC ALKALINITY A/Cl
0.1200
0.1250
0.1300
0.1350
DEPTH M
0
2000
4000
6000
8000
F
BURU BASIN
G
ARU BASIN
H
BANDA SEA
SPEC. ALKALINITY A/Cl

Fig. 11. F-H

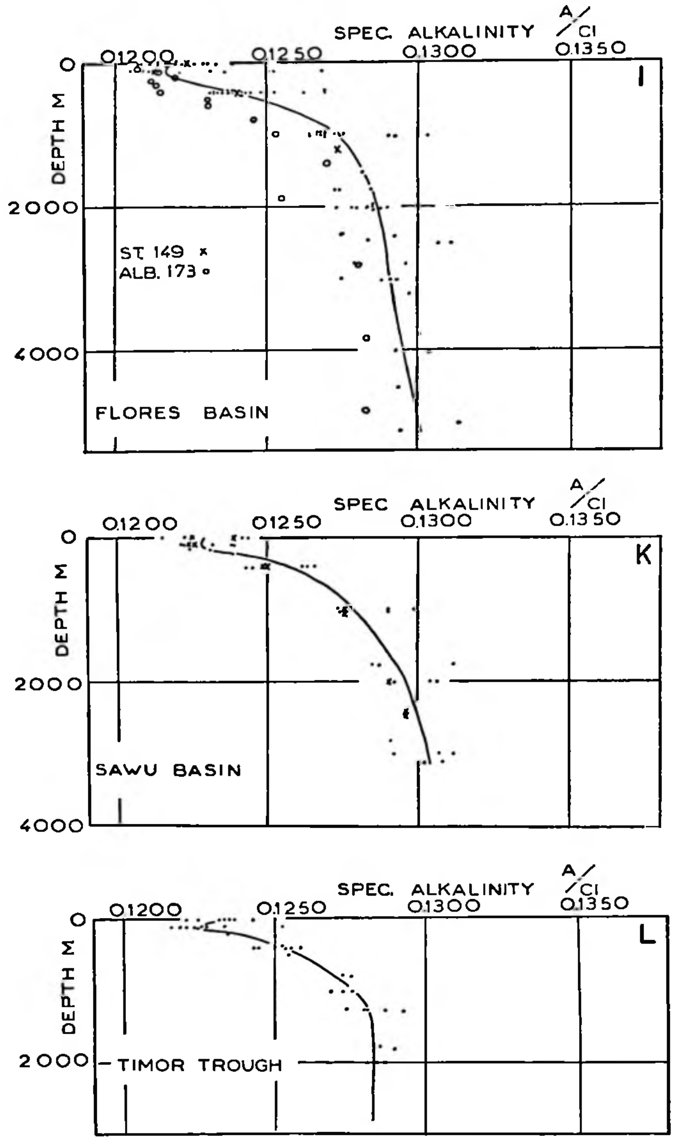


Fig. 11. I-L

b. Distribution of specific alkalinity. The main purpose of the following discussion will be a determination of the degree of saturation with calcium carbonate of the deep water of the basins. Fig. 12 represents the alkalinity-depth curves for most basins that are under the influence of the Pacific Ocean. The Sulu Sea data, collected with the first Celebes Sea series, have been corrected by subtracting 0.0030 specific alkalinity units from the original data of fig. 11 B. The Celebes Sea is represented only by the curve for the second Celebes Sea series. As a result of these corrections all curves of fig. 12 may in my mind be considered mutually comparable.

In all cases the specific alkalinity gradually increases with depth; this increase, for the oceans

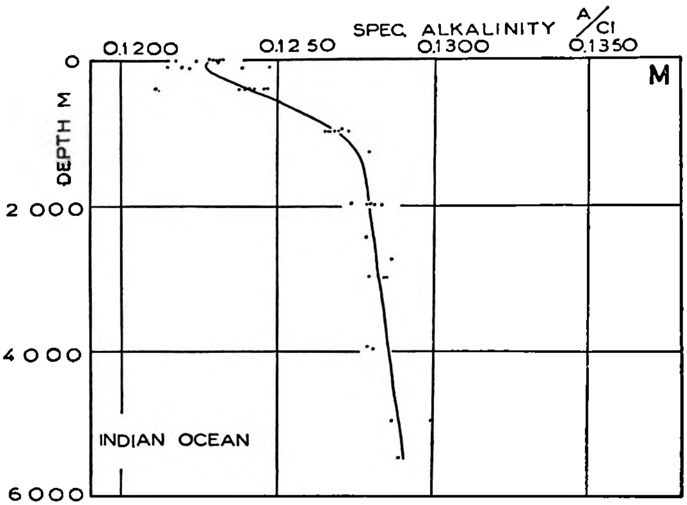


Fig. 11 M

first demonstrated by DITTMAR (1884), is due to plankton organisms with calcareous shells. These extract lime from the upper water layers, while, after death, they sink down to deeper water layers, where their skeletons, if not deposited permanently on the sea bottom, are dissolved.

The increase in specific alkalinity with depth also occurs in the open ocean. As a result, the alkalinity value in a basin below threshold depth depends in the first place on the depth of the sill over which the water enters: the higher the sill, the lower the specific alkalinity will be. Accordingly, the lowest specific alkalinity occurs in the Sulu Basin (saddle depth 400 m); next follows the Halmahera Basin (700 m), then the Celebes Basin (1400 m) and, finally, the basins supplied through the Strait of Lifomatola (1880 m). The basins of the Molucca Sea have not been added to fig. 12, since from these basins practically no measurements below sill depths are available.

Below the thresholds the specific alkalinity in the various basins, in general, does not remain constant with depth, but continues to increase towards the bottom; as a result the specific alkalinity in the deep water of a basin is mostly higher than that of the water entering over the threshold. This fact is better brought out, if also the vertical sections of specific alkalinity are considered (fig. 20 A-C). For example, the water flowing over the threshold of Lifomatola Strait has a specific alkalinity of 0.1270-0.1275 (section II). This value increases to about 0.1290 in the Buru and Banda Basins, to 0.1300 in the Weber Deep and the Flores Basin and to 0.1305 in the Sawu Basin. In the same way the specific alkalinity in the Celebes Basin increases from about 0.1260 above the sill to above 0.1270 in the deep water.

For the Sulu Basin the specific alkalinity of the inflowing water is not exactly known; according to the curve for the Pacific Ocean it may amount to 0.1230 (400 m). In the deep water of the Sulu Sea a specific alkalinity is found of about 0.1240. However, this estimate seems rather arbitrary, since only observations from two stations are available.

No increase of specific alkalinity was found in the Timor Trough and, probably, in the Halmahera Basin. In the latter case only one observation was available. The data in the deep water of the Aru Basin are erratic.

The increase of alkalinity below threshold depths proves that the deep water in most basins must be undersaturated with lime. The importance of this conclusion has been pointed out in the introduction. Furnished with this knowledge we may ask whether this undersaturation can also be proved by means of the values of the chemical properties determined. An accurate estimate of the saturation conditions on the basis of these properties—temperature, salinity, pH and alkalinity—has been made possible by the investigations of BUCH and WATTENBERG. Table 12 gives the percentage of saturation for the deepest part of the basins, calculated from the lime-solubility data of WATTENBERG (1933) and WATTENBERG and TIMMERMANN (1936).

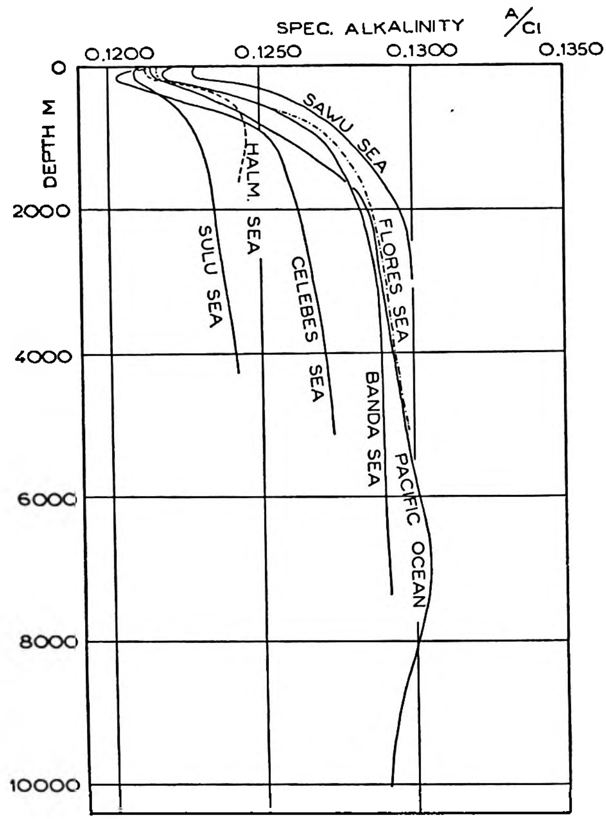


Fig. 12. Vertical distribution of specific alkalinity; corrected data (see text).

The lowest degree of saturation, viz. about 80%, is found in the deepest part of the Banda Sea (Weber Deep); the greatest oversaturation occurs in the Halmahera Basin, mainly owing to its high temperature. For the same reason also the Sulu Basin has a very high degree of oversaturation. The adjacent parts of the Pacific and the Indian Ocean are oversaturated down to great depths. Also the Timor Trench is oversaturated. The other basins are either undersaturated or close to saturation.

On the whole, the agreement between the degree of saturation and the solution of lime as evaluated from the vertical distribution of alkalinity is not unsatisfactory. Only in the case of the Sulu Basin does the alkalinity distribution seem to point to undersaturation, whereas according to the properties of the water this basin is oversaturated. This disagreement may be due to a lack of sufficient and reliable alkalinity data.

It will be clear from the above discussion that the conditions for the solution of lime are exceptionally favourable in most Indonesian basins, especially in those under the influence of the Pacific Ocean. The principal reason for this undersaturation is the comparatively low pH of the water, already mentioned before.

In connection with differences in the degree of saturation between the various basins the question arises whether these differences are reflected in the lime content of their sediments. At first sight, there seems to be a certain agreement; for example in the oversaturated Sulu Basin the average lime percentage of the deposits between 4000 and 5000 m amounts to 44%, whereas in the undersaturated Celebes and Banda Basins these percentages in the same depths are only 1 and 7.5% respectively (NEEB, 1943).

Table 12

Conditions for the solution of calcium carbonate in the deepest parts of the basins.

	Depth	pH in situ	Alkalinity, millaeq./L	Tempera-ture,in situ, °C	% Saturation [3]
Pacific Ocean	10500	7.66	2.48	2.4	89
	4000	7.75-7.80	2.48	1.6	108-113
Indian Ocean	6000 [1]	7.75-7.80	2.48	1.3	102-108
Sulu Basin	5580	7.72	2.37	10.5	121
Celebes Basin	6220	7.65	2.44	3.9	90.5
Makassar Trough	2540	7.73	2.42	3.6	97.5
Sangihe Trough	3820	7.70-7.75	2.44	2.4	96-106
Halmahera Basin	2039	7.80	2.40	7.8	130
Molucca Sea Basins [2]	4810	7.70	2.47	1.9	95
Buru Basin	5319	7.65	2.47	3.2	90
Banda Basins (Weber Deep)	7440	7.60	2.48	3.6	79
Flores Basin	5130	7.70	2.49	3.5	98
Sawu Basin	3470	7.65	2.49	3.4	91
Aru Basin	3680	7.70	2.47	4.0	100
Timor Trench	3310	7.75	2.47	2.8	107

[1]) Deepest "Snellius" observation [2]) Batjan Basin
[3]) Calculation according to WATTENBERG and TIMMERMANN (1936).

However, as regards the oversaturated basins one would expect that the lime content should be approximately as high as in an oversaturated part of the open ocean. If two equatorial areas are compared, there seems little reason to expect that the activity of lime-producing organisms would be very different in the two areas. Actually, however, the lime content of the oversaturated part of the equatorial Atlantic Ocean is down to a depth of about 4500 m above 60%, often even above 70%; in the case of the Sulu Sea the lime content does not rise above 50%, and in the other over-saturated basins the percentages are even much smaller.

Consequently, besides the properties of the water, another factor must play a rôle in decreasing the $CaCO_3$-percentage of the sediments and this factor is undoubtedly the high amount of terrestrial and volcanic matter which is deposited. Indeed by far the greater part of the area investigated is covered by terrigenous and volcanic muds (KUENEN, 1943 and NEEB, 1943) and the lime content of these muds (17%) is considerably lower than the average lime content of the muds and the pelagic sediments together (27%).

The relative proportions of the sedimentation of pelagic and terrigenous material have been determined by the above mentioned authors. They arrive at a rate of sedimentation of about 60 cm in 1000 years for terrigenous and volcanic material, excluding terrigenous lime. Beside this, a figure of 15 cm per 1000 years is assumed for the accumulation of lime, including lime from planktonic sources. These figures are much higher than those for the open ocean, where for the equatorial Atlantic Ocean SCHOTT (1939) assumes a total sedimentation of 1.8 cm in 1000 years and of 1.2 cm for calcium carbonate (see also KUENEN, 1950, pp. 376-386). There can be no doubt, therefore, that by far the greater part of the sedimentary material in the Indonesian Basins is of terrestrial origin.

It follows from this conclusion, that the average percentage of lime in the bottom material of the basins must be determined in the first place by the composition of the terrestrial and volcanic material. In this connection the dissolving action of the water will only play a secondary role by dissolving more lime in one basin than in another. The lime percentage in the basins oversaturated with lime will directly correspond with the lime percentage of the material settling into these basins.

The increase in specific alkalinity towards the bottom in the undersaturated basins allows an estimate of the amounts of calcium carbonate going into solution. If we take the basins of the

Banda Sea group as an example, we find that the water entering through Lifomatola Strait has above the sill a specific alkalinity of 0.1272, whereas the average lime content of the water inside the basins amounts to 0.1283 (compare p. 55). The specific alkalinity therefore increases by 0.0011. This increase corresponds with an alkalinity increase of 0.022 milliaequivalents per litre or 1.1 mg $CaCO_3$ per litre. The amount of water entering the basins has been estimated at 6300 km^3/year, (p. 23). Hence in one year about 7×10^{12} grams of lime would be dissolved in the Banda Sea and adjoining basins. The combined surface of the basins amounts to about $0.9 \times 10^{12}\,m^2$ (p. 23), so that per square metre every year about 8 grams of $CaCO_3$ would go into solution.

To decide whether at least the magnitude of this estimate would be acceptable, it should be compared with the amount of lime deposited permanently. KUENEN (1943) estimates this amount at 15 cm in 1000 years, or 10 grams of solid $CaCO_3$ per cm^2 in 1000 years, or 100 g/m^2/year. Consequently about 7% of the total amount of lime settling in the basins would go into solution. The remaining 93% would be buried before complete solution could take place, or be deposited in places where the water is saturated or oversaturated with lime.

The above relation between deposition and solution seems reasonable, although no great value should be attached to the percentages found. The principal conclusion must be that sedimentation and solution of lime in the Indonesian basins occur on a scale not directly comparable with that in the open ocean, where less than 10 grams of calcium carbonate are deposited on every square metre [1]).

The agreement between the lime content of the Indonesian sediments and those in the Pacific Ocean at 10-20° N mentioned before, must be considered as more or less accidental and does not indicate that the dissolving power of the deep water is in both cases the same.

c. *Specific alkalinity and water renewal of the basins.* From the data discussed in the previous chapter, and especially from table 12, it may be concluded that undersaturation of lime only occurs in the deeper parts of the basins. Solution of lime will, therefore, most probably take place only in water layers close to the bottom or even mainly from calcareous material deposited on the bottom (see also p. 58).

As a result, the increase in specific alkalinity is probably only brought about in the bottom water; this water mass is, so to speak, labelled by a higher specific alkalinity. The question may therefore be raised, whether, besides temperature, the distribution of specific alkalinity can give additional information about the way of water renewal of the basins. The following remarks mainly intend to draw attention to this question. They do not pretend to offer a definite solution of this problem.

The temperature distribution, together with the current measurements, shows that water enters a basin over its deepest threshold (VAN RIEL, 1934). This feeding current has a certain vertical extension; the deepest water, which has the lowest temperature, flows down along the inner slope of the sill and spreads over the floor of the basin, thus forming the bottom water. Higher layers of the feeding current have a slightly higher potential temperature and their water masses will flow down and spread along the potential isotherm corresponding with this temperature.

As a consequence of this inflow of water over the sill, the whole water mass inside a basin must gradually rise. Therefore, if no other forces were active, the isotherms would slowly be shifted upwards, and the more rapidly, the greater their distance from the bottom; at the same time the water mass in the basins would become gradually colder. The inflow of water would come to an end after the whole basin was filled homogeneously with the coldest, and therefore heaviest, water that can pass over the threshold.

Actually, however, processes are at work which keep the circulation going and keep the potential temperature inside the basin above that of the inflowing water. One of these processes will be vertical mixing of the water, caused by currents and tidal streams, through which heat is transported downwards. The other process will be the flow of heat from the earth, combined with convection currents. In principle, both processes cause an increase in potential temperature of the water inside the basins, accompanied by a downward shift of the potential isotherms, which in the steady state neutralizes

[1]) The supply of lime to the oceans from the land amounts to about 4 grams per square metre of the ocean floor per year (CLARKE, 1924), so that, if the lime concentration in the ocean water remains constant, the same amount should be deposited permanently.

the upward shift described above; in other words: the loss of heat caused by the replacement of basin water by colder water from outside is neutralized by a heat gain brought about by other processes.

The coldest inflowing water, which forms the bottom water of a basin, acquires near the bottom an elevated specific alkalinity, for the reasons mentioned above. In the state of equilibrium the amount of this bottom water rising upward will be the same through every level. However, with increasing distance from the bottom, the bottom water becomes mixed with increasing amounts of water of slightly higher temperatures and of the lower specific alkalinity occurring above the threshold. This intermixture causes the gradual decrease of specific alkalinity upward, which is shown by the vertical sections of fig. 20.

The specific alkalinity in the feeding current passing over the threshold is in all cases nearly constant throughout the whole water mass. Therefore, the differences in alkalinity between the bottom water, the water entering over the threshold and the water at a given depth are a quantitative measure of the percentage of bottom water present in the water mixture at this given depth. This percentage in its turn is a measure of residual upward water transport at this particular depth.

The above reasoning may be elucidated by a calculation for the complex of basins, embracing the Buru Basin, the Banda Basins, the Flores Basin and the Sawu Basin; for this complex the rate of water renewal is approximately known (p. 23). The vertical extension of the feeding current in the Lifomatola Strait may be estimated at 400 m (1480-1880 m). In the current the potential temperature decreases from 3.0° C at 1480 m to 2.5° C just above the threshold (fig. 17 B), whereas the specific alkalinity has the approximately constant value of 0.1272 (fig. 20 B).

The increase of specific alkalinity with depth inside the basins can be determined from figs. 20 B and 20 C. The course of the isopleths is rather complicated, especially in the east-west-section, so that the average specific alkalinity at a given depth can be determined only approximately. The average alkalinity values are entered in the table below for depth intervals of 500 metres. At 4200 m, the average depth of the complex of basins under consideration, the specific alkalinity is about 0.1297. From this level upward the specific alkalinity decreases down to 0.1283 at 2200 m, the upper boundary of the "homothermal" water mass.

The (unknown) quantity of water rising upward through the horizontal plane at 3700 m (specific alkalinity 0.1292) may be B m³ per year; the quantity passing through the horizontal plane at 3200 m (specific alkalinity 0.1289) may be $B + C$ m³ per year. The relation between the volumes B and C is:

$$0.1292\,B + 0.1272\,C = 0.1289\,(B + C)$$ [1]

The upward transport at 3200 m is therefore:

$$B + C = 1.18\,B.$$

In the same way we find at 2700 m (specific alkalinity 0.1286) an upward volume transport of 1.43 B and at 2200 m (specific alkalinity 0.1283) a transport of 1.82 B.

The value of B can be determined as follows. The total yearly inflow of water through the Strait of Lifomatola into the complex of basins amounts to 6300×10^9 m³ per year (p. 46). The same volume must leave the basins through the upper boundary of the water mass considered at 2200 m.

Therefore:

$$1.82\,B = 6300 \times 10^9 \text{ m}^3\text{/year}$$
$$B = 3465 \times 10^9 \text{ m}^3\text{/year}.$$

By entering this value in the above equations, the upward water transport through every layer can be determined; the result of this calculation is given in the following table.

Basins supplied through the Strait of Lifomatola.

Depth, m	A/Cl	Upward transport, $\times 10^9$ m³/year	Upward transport, m/year
4200	0.1297	—	—
3700	0.1292	3460	3.7
3200	0.1289	4070	4.5
2700	0.1286	4950	5.3
2200	0.1283	6300	6.7

[1] In the present calculation eddy diffusion of alkalinity has been neglected. Results of a refined analysis of the budgets of water, alkalinity and heat will be published elsewhere.

The upward transport, represented in the last column, has been calculated as follows. The time necessary for a complete renewal of the water in the basins has been estimated previously at 300 years. The average depth of the water mass under consideration is about 2000 m, so that the yearly rise of the upper boundary of the water mass should amount to 6.7 m/year. The upward rise at deeper levels can now be calculated from this value and the values for upward volume transport at these levels. For this calculation it has been assumed that the surface of the horizontal planes are at all depths about the same (900.000 km², p. 23).

The calculation leads to the result, that more than half of the water entering the basins flows down to the layers between 3700 m and the bottom (4200 m), with other words the bottom layers of the basins under consideration are very effectively ventilated. There is no reason why the same would not be valid for the other Indonesian basins. The strong ventilation of the deepest layers may be responsible for the absence of significant gradients of chemical values in the water layers close to the bottom (compare table 6, pp. 8-9).

As already discussed, the ventilation of the basins must be brought about by an increase of the temperature of the water in the basins. This increase in its turn must be due to heat flow from the earth's crust and to transport of heat from above by vertical mixing. The relative importance of the two contributions can be calculated as follows.

The average potential temperature of the water mass entering the basins amounts to 2.75° C (average potential temperature in the Lifomatola Strait below the isotherm of 3° C). The water rising upward through the upper boundary of the homothermal deep (2200 m) has a potential temperature of 3.0° C. Hence, the total increase of temperature of the deep water, when leaving the basin, amounts to about 0.25° C.

The heat transport through the earth's crust probably amounts to about 0.5×10^6 cal/m² per year (Kuenen, 1943, Bullard, e.o., 1954, 1956). The upward transport of water at 2200 m amounts to 6.7 m per year. The temperature increase of this water, due to heat from the earth, must therefore amount to $\frac{0.5 \times 10^6}{6.7 \times 10^6} = 0.07°$ C.

According to this calculation, the temperature increase caused by heat diffusion from above must amount to 0.25 — 0.07 = 0.18° C. Hence the warming up of the deep water must be caused for about one third by heat from the earth's crust and for two third by heat transport from above.

The same calculation can also be carried out for water layers below 2200 m. It appears that the influence of heat transport from above decreases with depth, whereas that of heat from below increases with depth. As a result, at 3700 m about nine tenths of the temperature increase appear to be due to heat from the earth's crust. Most probably the ventilation of the deepest layers in the basins is therefore mainly generated by the flow of heat from the earth.

It has already been observed, that the distribution of specific alkalinity in the Banda Sea and adjacent areas is more complicated than indicated by the north-south section II (fig. 20 B) alone. For this reason alone the above calculation must be considered a very approximative one. Along section IV (fig. 20 C), which crosses the region from west to east, the course of the alkalinity isopleths is as follows: from station 246, where the two sections II and IV intersect, the isopleths slope upward both to the left and to the right. The slope is especially well marked in the Flores Basin, where the isopleths rise from the entrance between the stations 194 and 197 toward the closed western end. Roughly speaking, this course forms an indication of the differences in "age" occurring in the water masses of this basin; the oldest water evidently occurs in its western corner, where the specific alkalinity attains exceptionally high values. The potential isotherms of this basin (fig. 17 D) slope in a direction opposite to that of the alkalinity lines.

Also in the eastern part of section IV the latter isopleths slope upward. In this area, besides the age of the water, the very considerable undersaturation of the water in the Weber Deep may play an additional role in causing high alkalinity values.

d. *Comparison with other areas.* The observations of specific alkalinity in the parts of the oceans adjacent to the Indonesian Archipelago, together with the results of measurements in the equatorial Atlantic Ocean (Wattenberg, 1933), give rise to some remarks concerning the cycle of calcium carbonate in the oceans.

Fig. 13 represents alkalinity-depth curves of the above-mentioned areas. It appears that in the deep water the specific alkalinity increases in the sequence Atlantic Ocean-Indian Ocean-Pacific Ocean, a fact observed previously by the expeditions of the "Carnegie" (REVELLE, 1936) and "Albatross" (KOCZY, 1956). Above 1600 m the part of the Indian Ocean studied has a higher specific alkalinity than the Pacific Ocean. This may be due to an increase in specific alkalinity in the water passing through the Archipelago from one ocean to the other. This increase may be caused by vertical mixing, a process which will lead to an enrichment in lime of the upper water layers by the addition of lime dissolved in the basins.

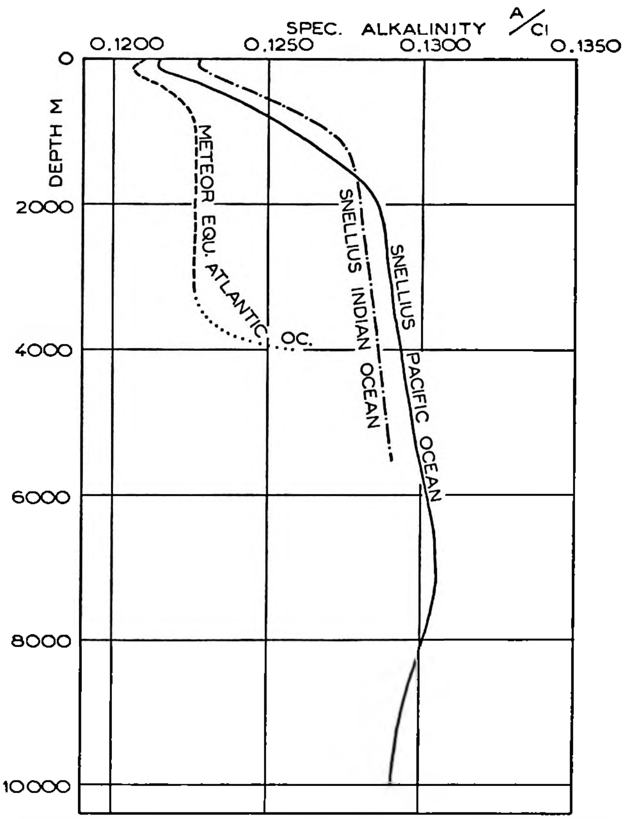


Fig. 13. Vertical distribution of specific alkalinity in the equatorial Atlantic Ocean, after Wattenberg, 1933, and in the Pacific and Indian Ocean, according to the "Snellius" observations.

The differences in specific alkalinity between the deep waters of the three oceans are caused by differences in the deep water circulation. In the Atlantic Ocean deep water is formed in the north; this North Atlantic Deep Water penetrates towards the south across the Equator. Counter to this transport surface and subsurface water flows towards the north. These water masses have a comparatively low specific alkalinity, and as the net transport of calcium across the Equator must be zero, the lime content of the North Atlantic Deep Water must also be low. As a result the North and the Equatorial Atlantic Ocean have a comparatively low lime content.

Another source of deep and bottom water is located in the Antarctic Ocean. In the Atlantic Ocean this water flows beneath the North Atlantic Deep Water to the north. Much of it returns to the south after mixing with the south-moving deep water; this fact favours a rapid renewal also of the bottom water in this ocean.

In the Indian and the Pacific Oceans no deep water is formed in the north. The deep water

flowing in from Antarctic regions only, the penetration of deep and bottom water to the north is very slow, since a water circulation by which it can rapidly return to the south is absent. Therefore, the deep and bottom water in the northern Pacific and the Indian Ocean is much older than in the Atlantic Ocean. The older the water mass, the longer will be the period in which enrichment with lime from sinking organisms may be assumed to have taken place. As a result the Pacific and the Indian Ocean have a much higher lime content than the Atlantic Ocean.

The same reasoning also holds for other substances which as a result of biological or other processes sink down from the surface layers into deep water. Actually, all these substances must show a similar distribution over the oceans. The best known example in this connection is the distribution of phosphate.

A factor, however, which must play an important rôle in causing concentration differences between the deep and surface layers and between one ocean and another, is the depth in which the various substances are brought into solution. To illustrate this we will compare the accumulation in the deep water of dissolved calcium carbonate and phosphate. Regarding phosphate it is known, that most of this compound, probably even 90% (RILEY, 1951), is regenerated from organic matter already in the upper 1000 metres. The depth in which calcium carbonate is dissolved needs some further consideration.

KUENEN (1950, p. 354) is inclined to believe that by far the greatest amount of lime is dissolved on the sea bottom and not during the period of sinking, since this period is so much shorter than the time during which the calcareous skeletons lie unprotected on the bottom. Moreover, the skeletons are originally protected against solution by organic material strongly resistant against decomposition. The cover will probably not be decomposed before the skeletons reach the bottom.—To these arguments may be added the fact that the water column through which the particles settle is always over a large part and sometimes over its whole length oversaturated with lime.

This point of view is supported by the distribution of the specific alkalinity in the Atlantic Ocean, where the largest increase occurs close to the bottom (fig. 13). Also in the Indian Ocean an increase close to the bottom has been observed (KOCZY, 1956), although this increase is not shown by the "Snellius" observations.

In the Pacific Ocean, on the contrary, a decrease of alkalinity is found below certain depths and no increase occurs near the bottom (KOCZY, 1956). The "Snellius"-observations show an alkalinity maximum at 7000 m. These facts find a good explanation if we assume that at least part of the lime is dissolved in layers at some distance from the bottom. Two factors may promote a more rapid solution of lime in the Pacific Ocean: the comparatively high degree of undersaturation and the comparatively great depth.

However, even if in the Pacific Ocean solution of calcium carbonate takes place before the lime particles reach the bottom, there can be no doubt that this process only begins in depths far below 4000 m (table 12). Taking all things together, we conclude that the solution of lime only occurs in the deepest water layers of all oceans.

The above discussion shows that there is a considerable difference in depth of "solution" of lime and phosphorus. Therefore, the accumulation of lime in the deep and bottom water of the oceans must be much more effective than that of phosphate; in other words: the sinking of a certain amount of lime from the surface layers leads to a greater increase of lime in the deep water than the sinking of an equal amount of phosphate. As a result, the increase of lime with increasing age of the deep and bottom water will be more rapid than the corresponding increase in phosphate. This implies, that the difference in concentration between the surface and the deep water of the oceans will be greater for calcium carbonate than for phosphate (all this to be taken in relation to the amounts of the two compounds settling down from the surface layers).

The following calculation may elucidate this consideration. If we take all available data together, we find that the specific alkalinity of the surface water of the oceans on an average amounts to 0.1215 and of the deep water (4000 m) to 0.1270; the difference, 0.0055, corresponds with about 5.5 mg $CaCO_3$ per litre. The corresponding phosphate values, according to measurements by several expeditions, are probably about 0.3 and 2.5 μg-at/l; the phosphate difference between the deep and surface water therefore amounts to about 2.2 μg-at/l or 0.066 mg phosphorus per litre. Hence, about 85 mg of lime are accumulated against 1 mg phosphorus. The relation 85 : 1 is certainly out of

proportion in comparison with the composition of the average plankton population sinking down from the surface layers, in which the lime content probably amounts to no more than a few times that of the phosphorus content. Evidently the error in this calculation is, that no account has been taken of the fact that phosphorus, which is for the greater part regenerated in the upper water layers, passes many times through the cycle of production and decomposition of organic matter, before reaching the deep water, while lime, after being included into skeleton material, at once sinks down to the greatest depths.

A factor which influences the above calculation, but which for the sake of simplicity has so far been left out of account, is the deposition of lime on the sea bottom. In areas, where the water column is oversaturated down to the bottom, the lime skeletons will not return into solution and be deposited permanently; this fraction will therefore not participate in the enrichment with calcium carbonate of the deep and bottom water. As a result, the actual relation between the amounts of calcium and phosphorus sinking down into deep water must be even less favourable for phosphorus than the relation determined above.

The most conspicuous difference exists, in this respect, between the Atlantic Ocean and the equatorial and northern Pacific Ocean. In the former area, which is in general oversaturated, less lime will return into solution than in the latter, undersaturated area. This will cause an extra difference in dissolved lime content between the two oceans, apart from the difference caused by the deep water circulation. Phosphate, on the contrary, is probably regenerated in both oceans at the same rate.

The following calculation illustrates this point. The specific alkalinity of the deep water in the equatorial Atlantic Ocean amounts to about 0.1230; the corresponding phosphate content is about 1.6 μg-at/l (WATTENBERG, 1938). In the deep water of the Pacific Ocean near the Philippines the specific alkalinity is on an average 0.1290 and the phosphate content, according to the "Dana"-observations (fig. 14), amounts to 3 μg-at/l. From these values the difference in lime content between the deep water of the two oceans can be calculated to amount to 7 mg of $CaCO_3$ per litre, whereas the difference in phosphorus content equals 0.048 mg of phosphorus per litre. The accumulation in the deep water of 7 mg of lime against 0.048 mg of phosphorus should be compared with the difference between the deep and surface water, calculated above, of 5.5 mg of lime against 0.066 mg of phosphorus. It appears that lime is about twice as well accumulated in the Pacific Ocean as in the Atlantic Ocean.

Summarizing, we arrive at the conclusion that the distribution in the oceans of chemical substances participating in the cycle of organic matter, is in the first place determined by the character of the oceanic deep water circulation. In addition other factors, such as the depth of solution and regeneration and permanent deposition on the sea floor seem to play a rôle in determining quantitative differences.

e. *The minimum in specific alkalinity at a depth of 100 m.* An aspect of the alkalinity distribution which finally needs our attention is the existence in many stations of a minimum in specific alkalinity at a depth of about 100 m. Of the 94 stations with suitable measurements (i.e. observations at 0,100 and 400 m), 61 stations show this minimum clearly, at 10 stations no difference between 0 and 100 m is observed and at 23 stations the minimum is absent (table 15). Of the latter stations no less than 7 are located in the Flores Sea, where in the majority of cases the minimum is lacking; the other 16 are scattered over the whole area of investigation.

We must conclude, therefore, that the alkalinity minimum is a feature of the whole area and not restricted to a certain part of it. Starting from this point of view, the question of the origin of this minimum may be considered. There are two possible explanations, both already put forward by other authors in connection with similar minima in the open oceans (WATTENBERG, 1933, KOZCY, 1956). One explanation is, that the optimum conditions for the growth of the calcareous plankton are found at the depth of the minimum. The other explanation is, that the minimum is connected with the salinity maxima at 100-200 m. In the latter case the low specific alkalinity is caused by the consumption of lime in the surface water of the places where the salinity maxima are formed.

If the latter explanation should be correct, a close relation between the distribution of the alkalinity minimum in the Indonesian seas and the salinity maximum penetrating into the Indonesian waters from the Pacific Ocean were to be expected. The observations show, however, that

this relation does not exist. Moreover, the actual depth of the alkalinity minimum is probably less than 100 m (Koczy, 1956). Therefore the first explanation seems more probable.

C. *Phosphate.* The number of phosphate determinations carried out falls considerably short of the number relating to the other chemical factors. Moreover, the estimations were made only during the second part of the expedition and they cover, therefore, only part of the area investigated. Another serious drawback, already mentioned in chapter II, is caused by the fact that the phosphate values found are too low, so that they cannot be directly compared with those of other expeditions. These shortcomings seriously impair the value of the determinations.

Some conclusions can nevertheless be drawn from the available data, especially if the observations of the "Dana" expedition are also considered. Figs. 14 and 15 represent average depth-phosphate curves for various basins; fig. 14 is constructed by means of the "Snellius" data and fig. 15 represents the "Dana" observations. It appears from these figures that the phosphate content of the Sulu Basin is exceptionally low. This fact can easily be explained as the result of the shallowness of the passage connecting the Sulu Basin with the China Sea. The figures further show that below the threshold the phosphate content remains constant with depth or shows only a very slight increase. This is in accordance with the very slight decrease in oxygen in the deep water.

D. *Summary of chemical investigations.*

The numerous water samples collected and analysed by the "Snellius"-Expedition provide a fairly complete picture of the distribution of oxygen, pH and alkalinity in the East Indonesian waters. The phosphate data are less satisfactory in this respect. In considering the distribution of chemical values a distinction must be made between the water masses above the thresholds separating the deep sea basins and the water masses filling these basins. Those above the threshold will be considered first.

In the case of oxygen, stress has been laid on the observation that in the two main water masses present in the Pacific Ocean between Mindanao and New Guinea (compare summary of water movement, p. 27) oxygen minima are present at intermediate depths. The northern minimum, however, is located in a higher water level (400 m) than the southern minimum (1000 m); moreover, the oxygen content of the northern water is lower than that of the southern water, except in the water levels above 200 m, where the reverse appears to be the case. The oxygen distribution in the Indonesian Archipelago is mainly determined by these differences; yet, certain changes occur in the water masses during their flow through the area, which cause the picture to be somewhat more complicated. One of these changes is the gradual fading of the minimum in the southern waters, which enters between Celebes and Halmahera; the northern minimum, which follows a course through the Celebes Sea, the Strait of Makassar and the Flores Sea, remains well pronounced as far as the Indian Ocean. The other change is caused by the loss of oxygen during the transport of the water from north to south; this loss becomes more conspicious, when the water flows more slowly and is especially felt, therefore, in the eastern Banda and Timor Seas and in the Arafura Sea. The decrease of oxygen, together with approximative estimates of the rate of flow of the water, produced some insight into the consumption of oxygen in the different water levels.

The distribution of pH follows, generally speaking, that of oxygen in this sense that a low oxygen content is accompanied by a low pH. The alkalinity data are not sufficiently accurate to allow such a sharp distinction to be made between the concentrations in the different water masses as in the case of the above-mentioned values; it seems fairly certain, however, that the specific alkalinity increases from the Pacific Ocean towards the Indian Ocean, which may be due to the solution of lime in the deep water of the basins, as will be mentioned below.

As regards the properties of the water masses in the basins, Van Riel has shown that their temperature, salinity and oxygen values are mainly determined by the depths of the sills. This conclusion was confirmed in detail for oxygen and was shown to hold as well for pH, specific alkalinity and phosphate. Since most of the sills are in depths not far from the depths of the oxygen and pH minima, the oxygen and pH values in the basins are generally low if compared with those at corresponding depths in the oceans. Specific alkalinity in the oceans mostly increases regularly with depth, so that the specific alkalinity of the basins was found to be higher with increasing depth of the threshold.

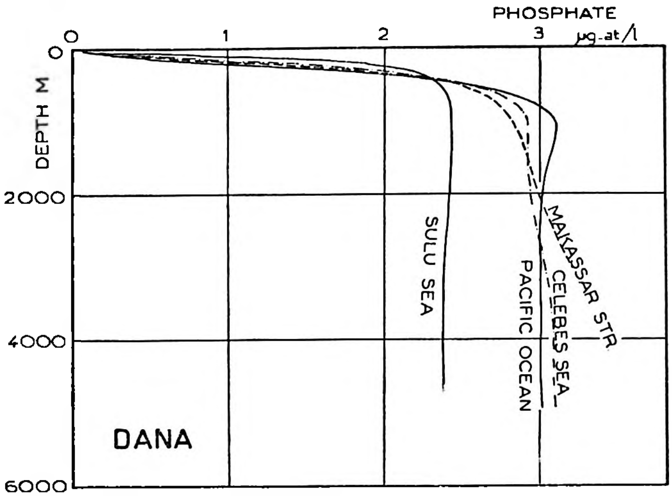


Fig 14. Vertical distribution of phosphate in the Pacific Ocean and in the Indonesian basins according to the observations of the "Dana"-expedition (1928-1930); the phosphate values have been corrected for salt error.

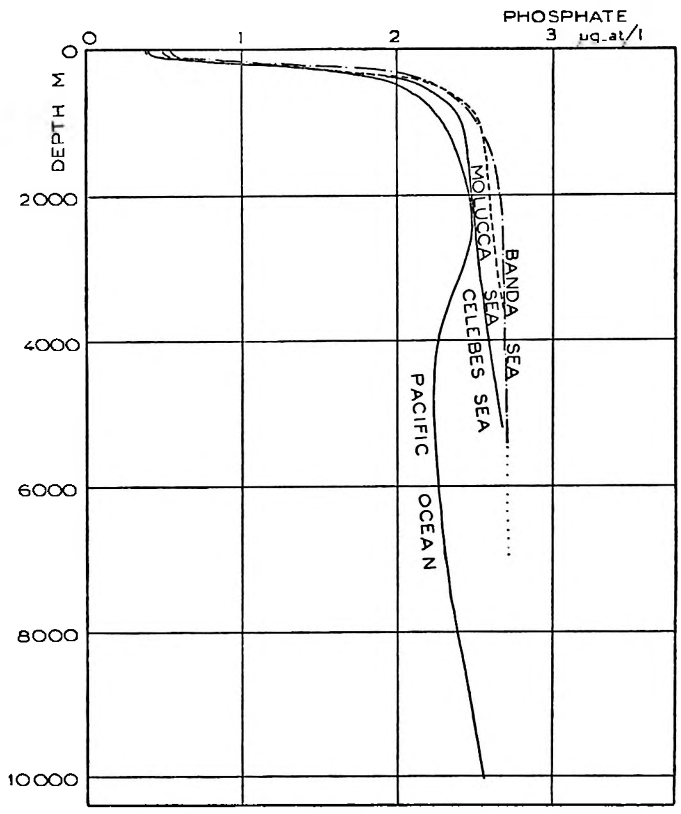


Fig. 15. Vertical distribution of phosphate in the Pacific Ocean and in the Indonesian basins according to the observations of the "Snellius"-expedition; the phosphate values have been corrected for salt error.

Inside the basins, slight changes in the initial concentrations of the various properties appeared to occur. Oxygen and pH showed a small decrease, compared with their values in the inflowing water; this is especially the case in the Flores and Sawu Basins and must be due to oxygen consumption and carbon dioxyde production in situ by biological processes. The same holds for the small increase found for phosphate.

In many basins a slight increase below threshold depth was also found for specific alkalinity; this pointed to solution of lime in the deep water, so that these basins must in this respect be undersaturated. This conclusion was confirmed by a calculation for various basins of the degree of saturation for lime; this appeared to vary between 130% (oversaturation) in the deepest part of the Halmahera Basin to 79% (undersaturation) in the Weber Deep. Taking all data together, the conditions for the solution of lime in the deep water of the basins appeared to be very favourable, a fact that must in the first place be attributed to the comparatively low pH of the water masses. For the basins which receive their water through the Strait of Lifomatola (Banda Basin, etc.), the amount of lime going into solution could roughly be estimated. By also taking the sedimentary data of KUENEN and NEEB into account, the conclusion could be drawn that both sedimentation and solution of lime occur in the basins on a much larger scale than in the oceans.

The distribution of specific alkalinity in the water masses below the thresholds further induced considerations about the process of water renewal in the basins, from which it was concluded, in agreement with certain suppositions made by KUENEN, that in this connection heat flow from the earth's crust must play an important role.

Finally, the distribution of specific alkalinity in the Indonesian seas was compared with those in the oceans; in this connection attention was given to the process of accumulation of lime in the deep and bottom water of the oceans, as compared with phosphate.

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Fig. 16 A. Vertical distribution of salinity; section I; compare fig. 2.

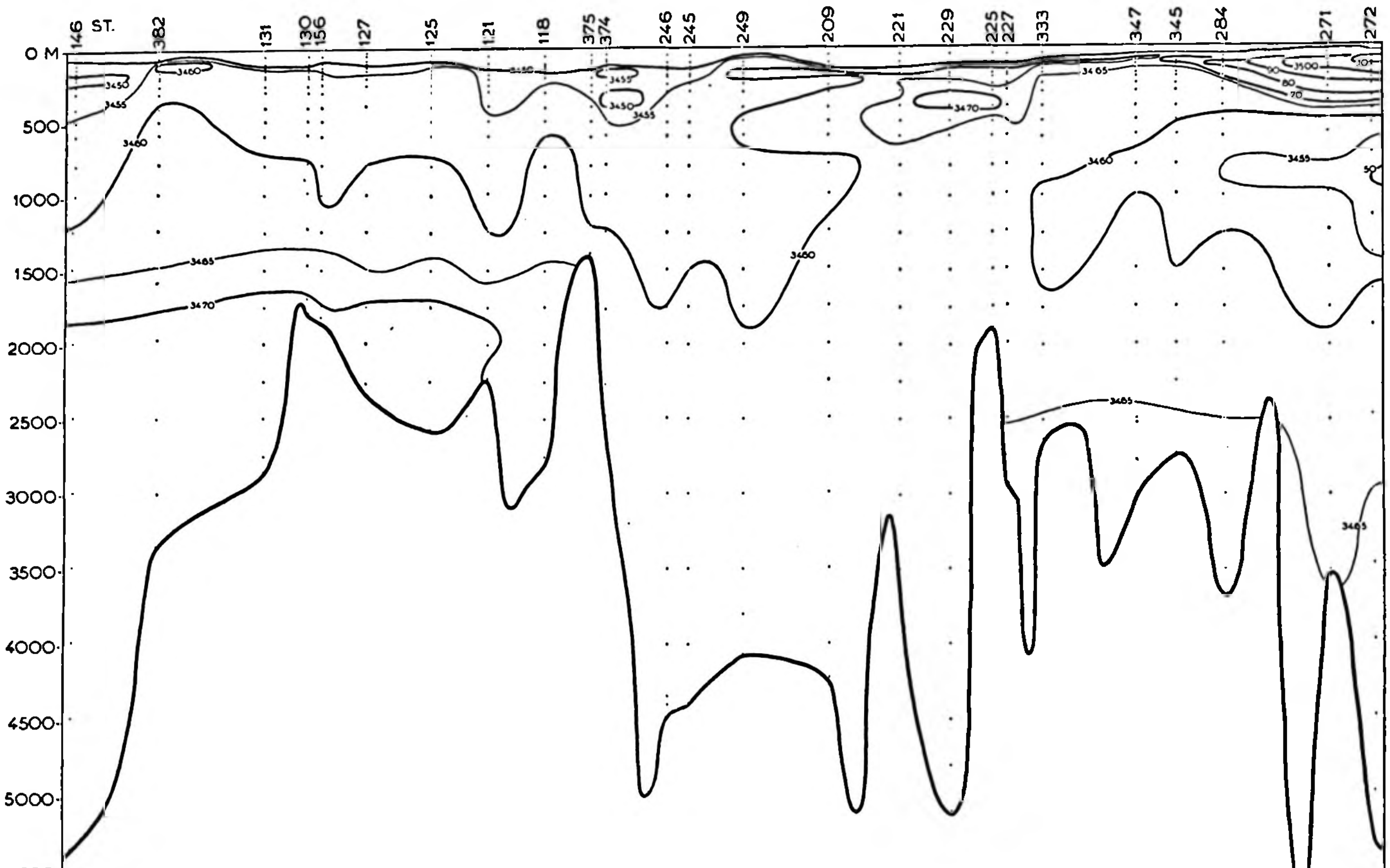

Fig. 16 B. Vertical distribution of salinity; section II; compare fig. 2.

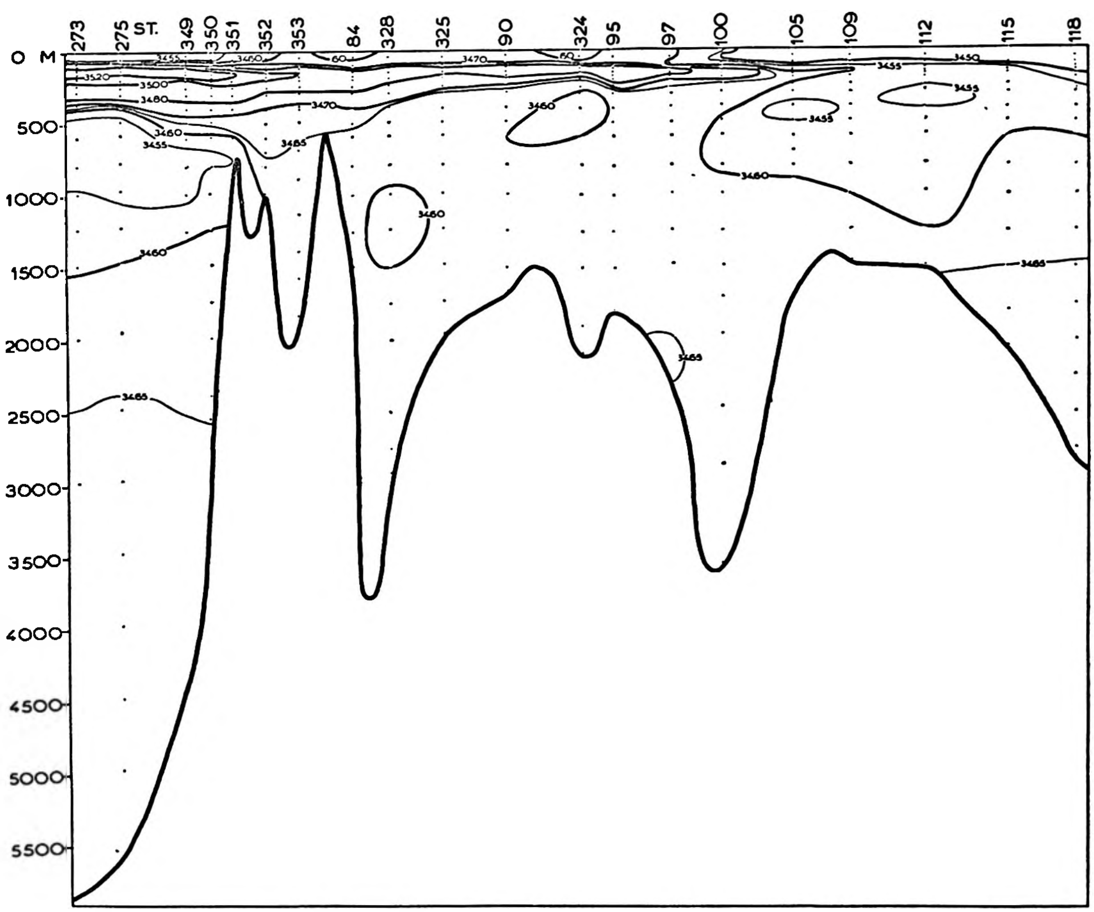

Fig. 16 C. Vertical distribution of salinity; section III; compare fig. 2.

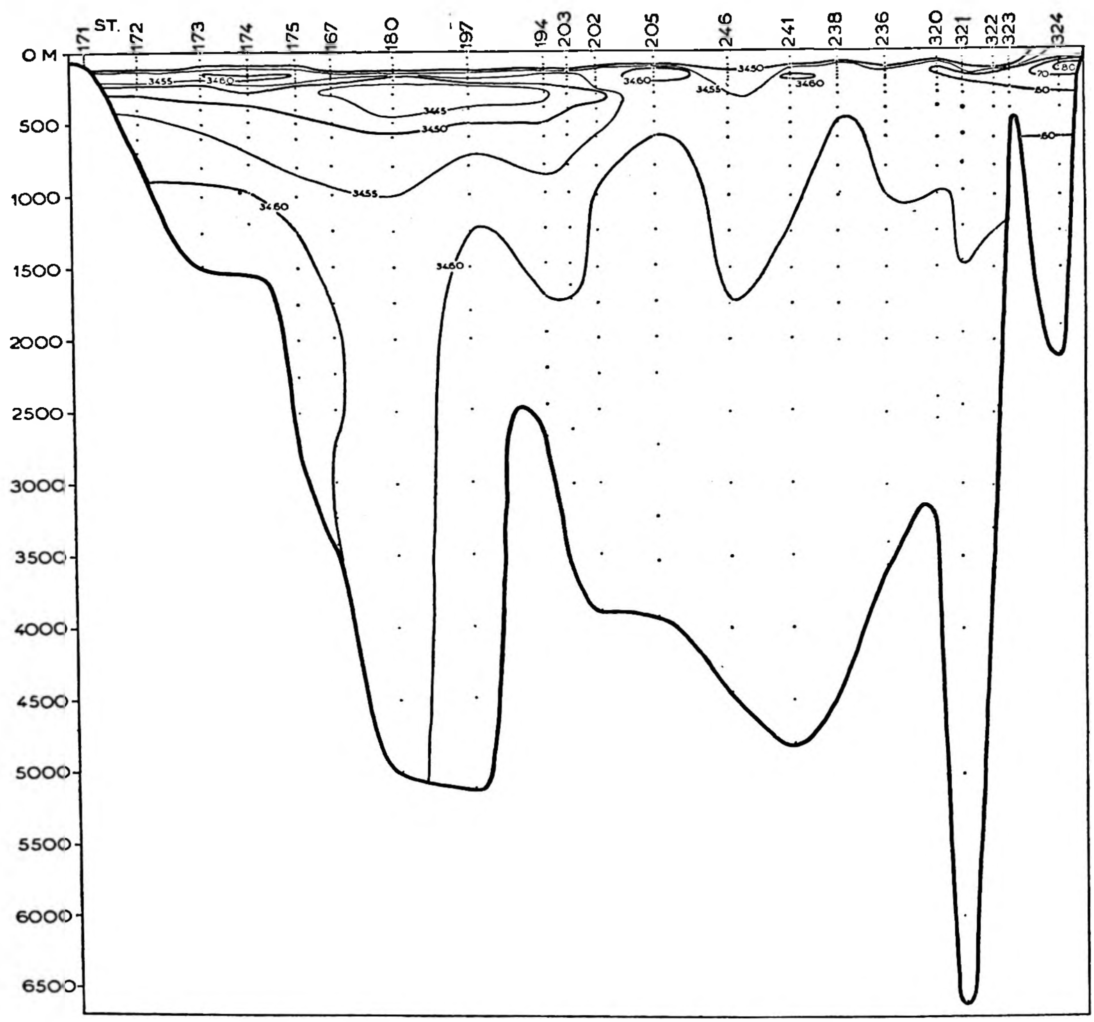

Fig. 16 D. Vertical distribution of salinity; section IV; compare fig. 2.

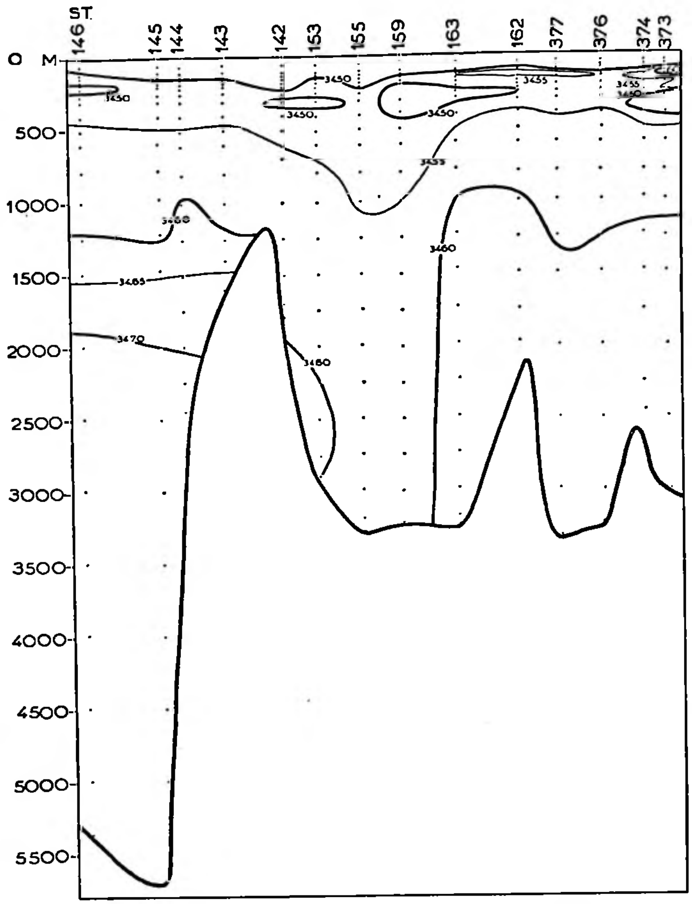

Fig. 16 E. Vertical distribution of salinity; section V; compare fig. 2.

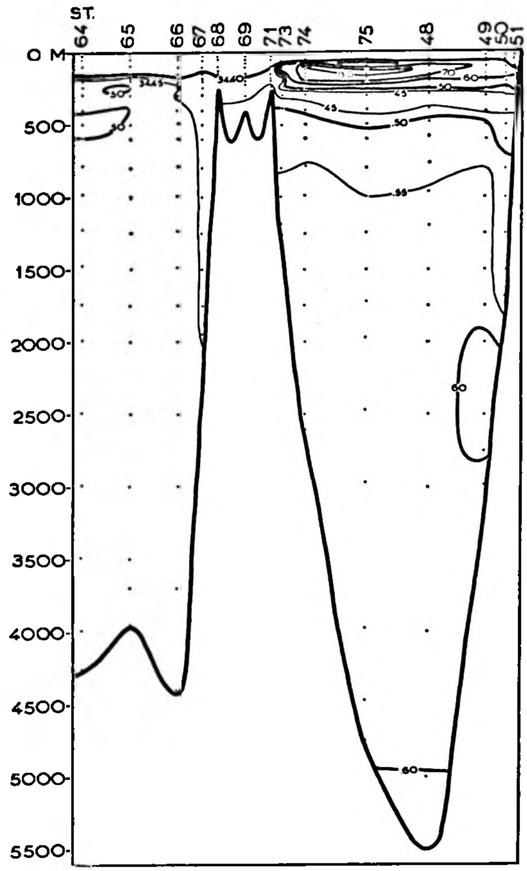

Fig. 16 F. Vertical distribution of salinity; section VI; compare fig. 2.

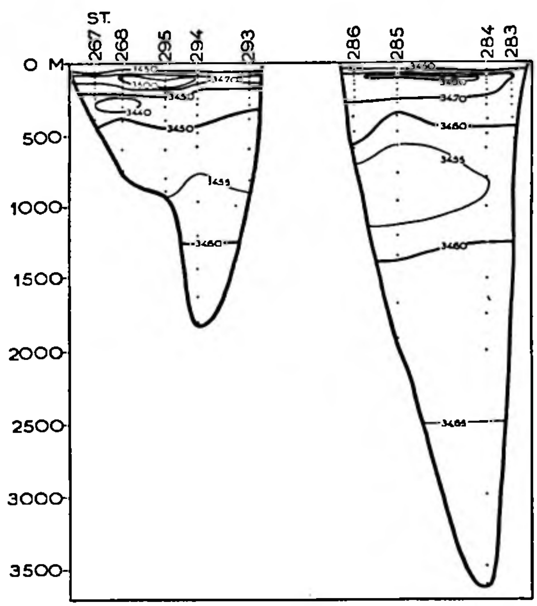

Fig. 16 G. Vertical distribution of salinity; section VII; compare fig. 2.

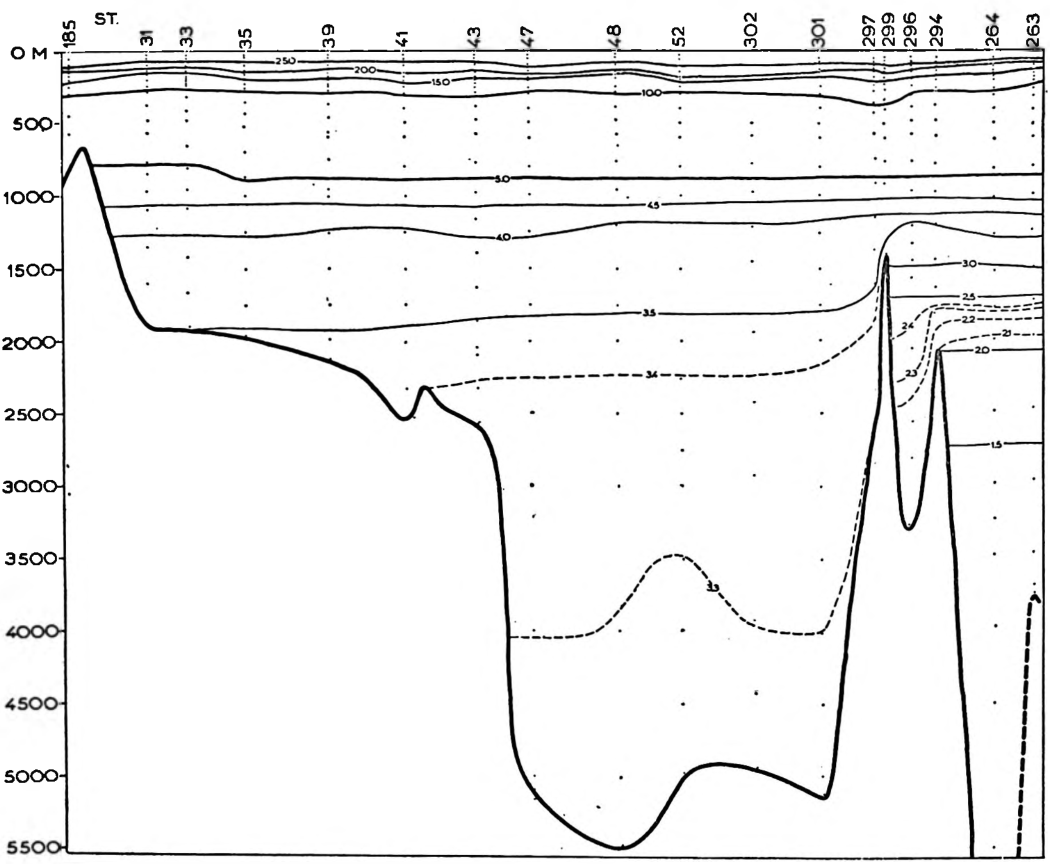

Fig. 17 A. Vertical distribution of potential temperature; section I; compare fig. 2.

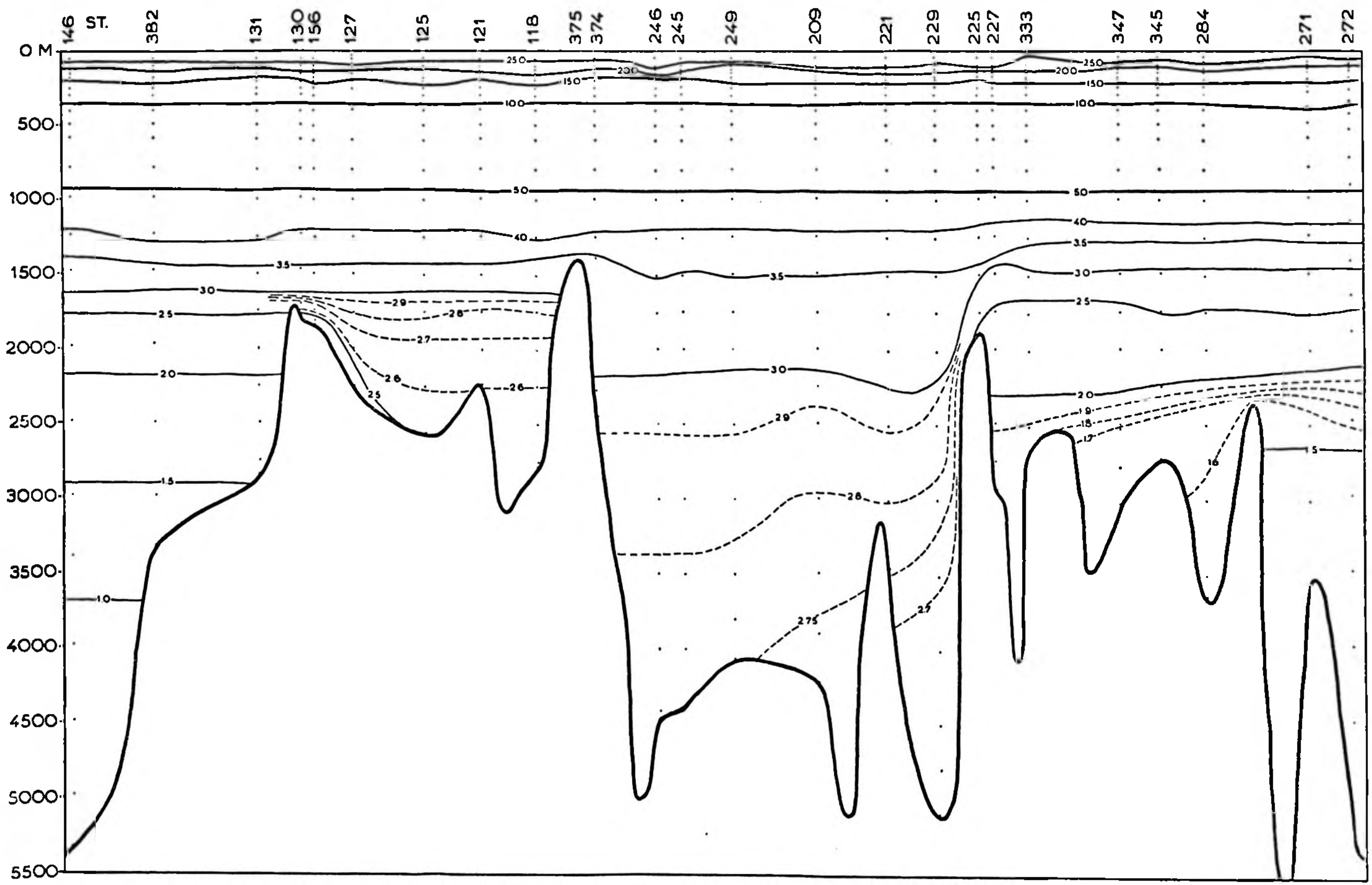

Fig. 17 B. Vertical distribution of potential temperature; section II; compare fig. 2.

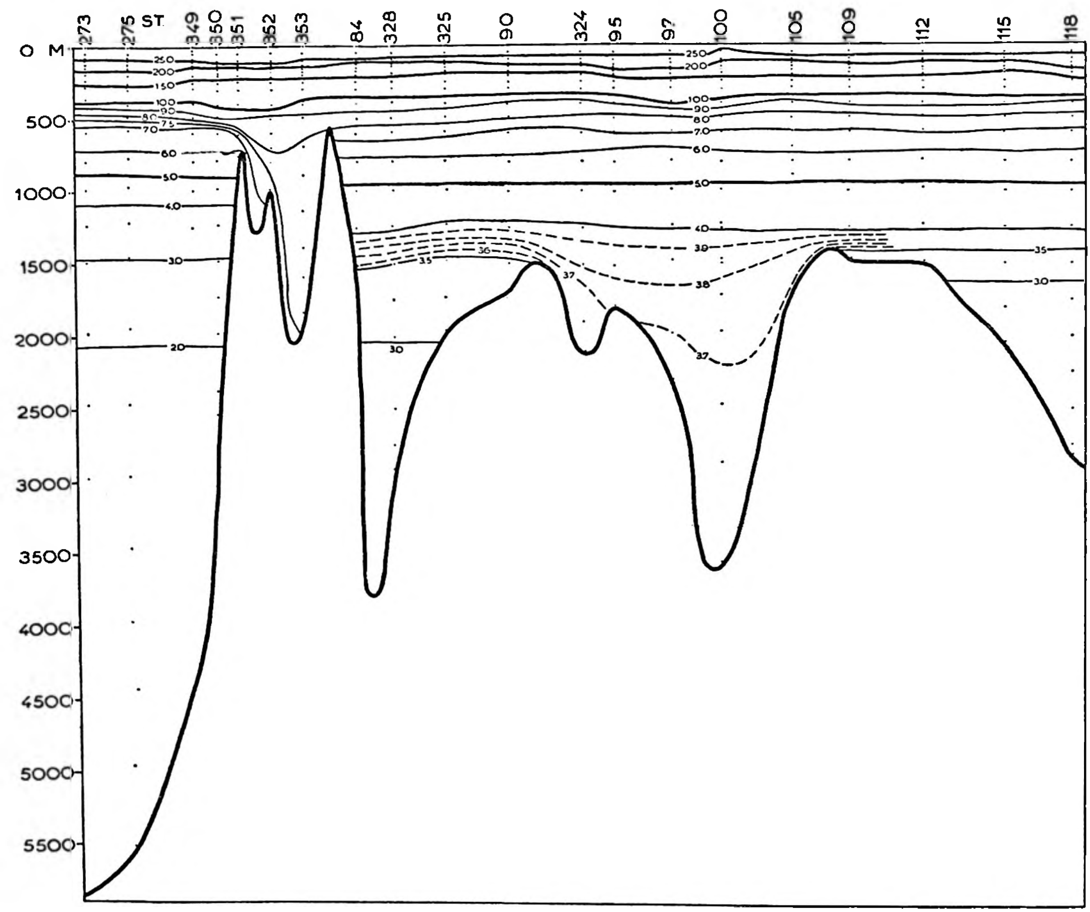

Fig. 17 C. Vertical distribution of potential temperature; section III; compare fig. 2.

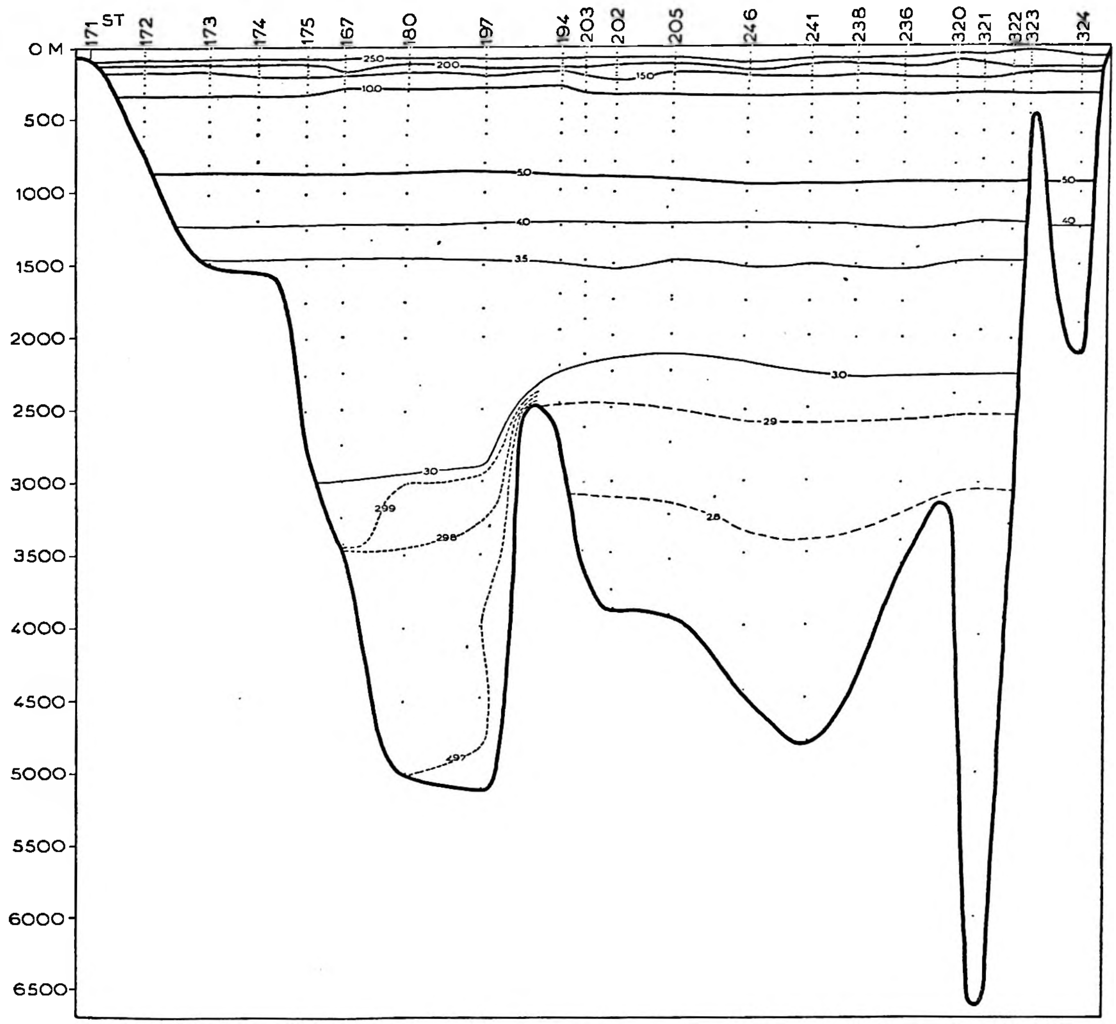

Fig. 17 D. Vertical distribution of potential temperature; section IV; compare fig. 2.

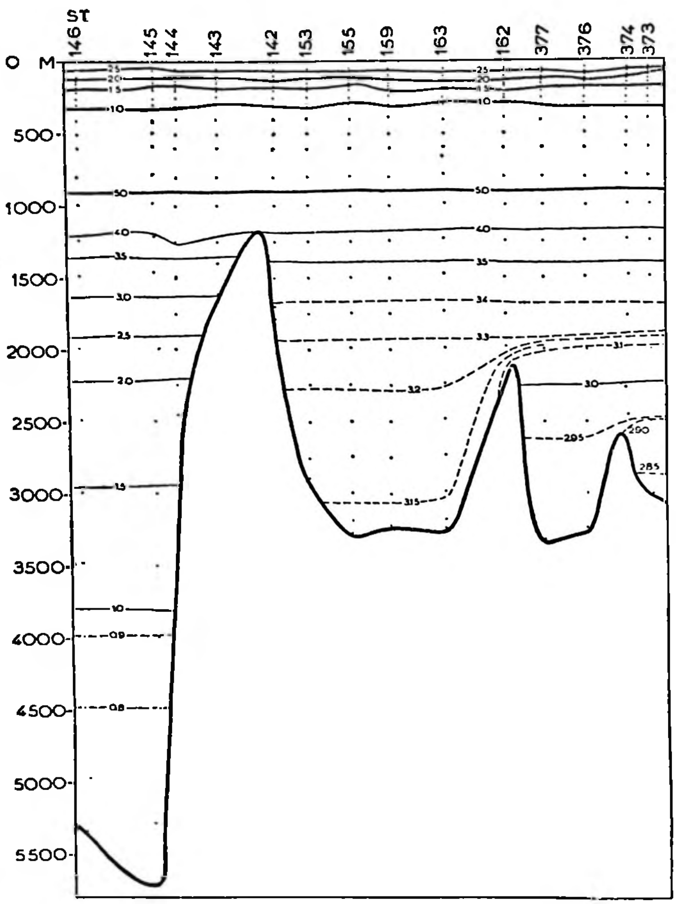

Fig. 17 E. Vertical distribution of potential temperature: section V; compare fig. 2.

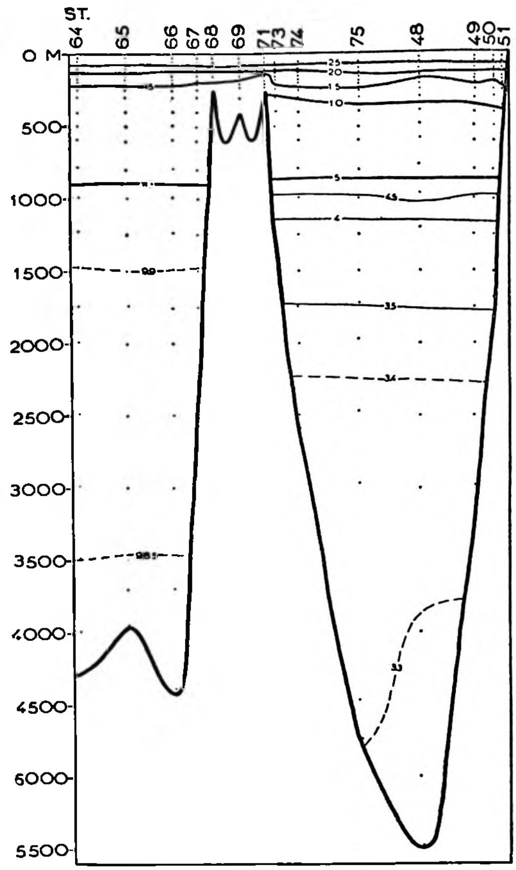

Fig. 17 F. Vertical distribution of potential temperature; section VI; compare fig. 2.

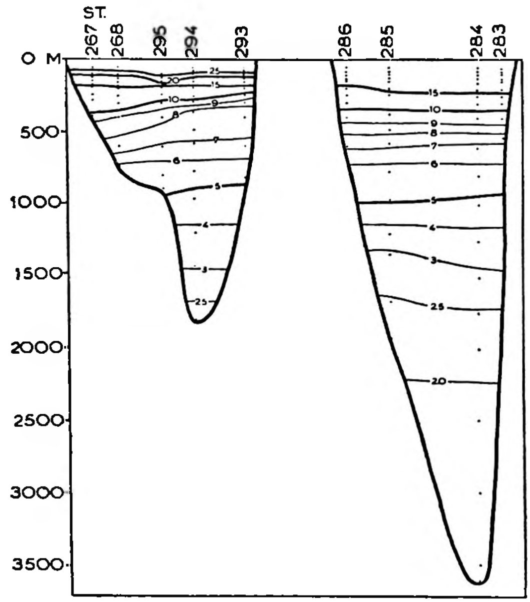

Fig. 17 G. Vertical distribution of potential temperature; section VII; compare fig. 2.

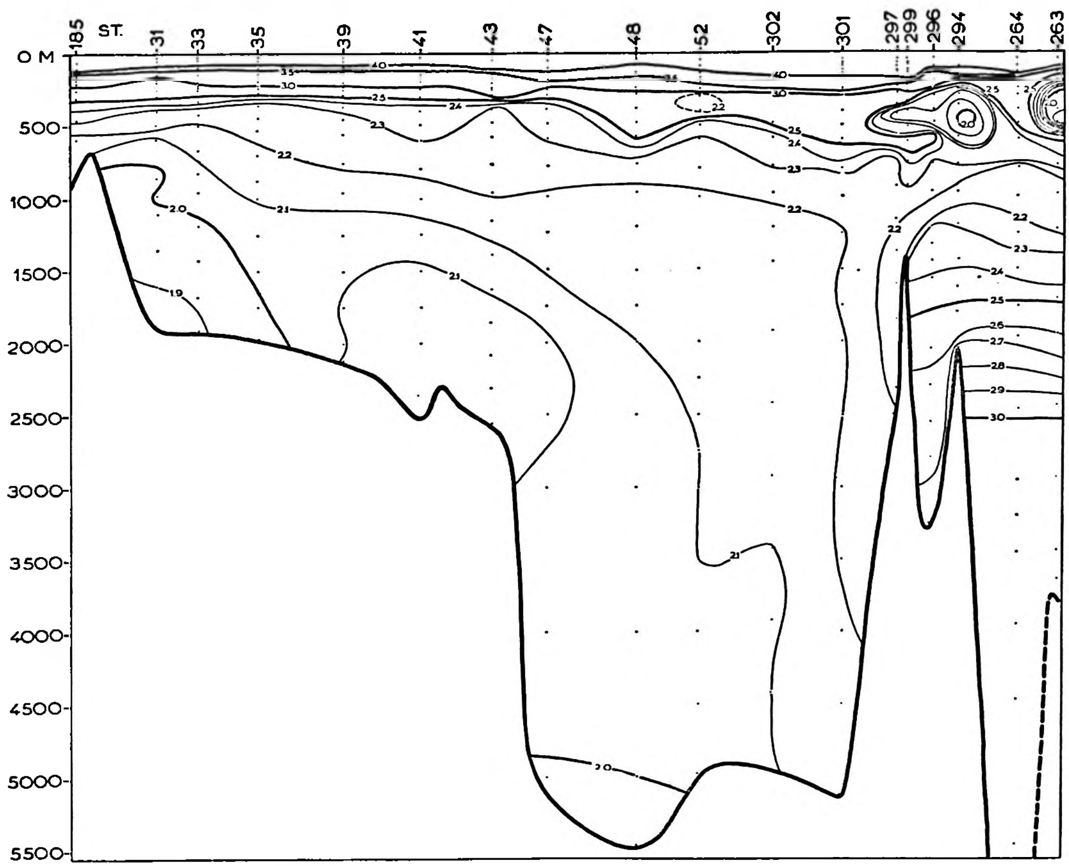

Fig. 18 A. Vertical distribution of oxygen; section I; compare fig. 2.

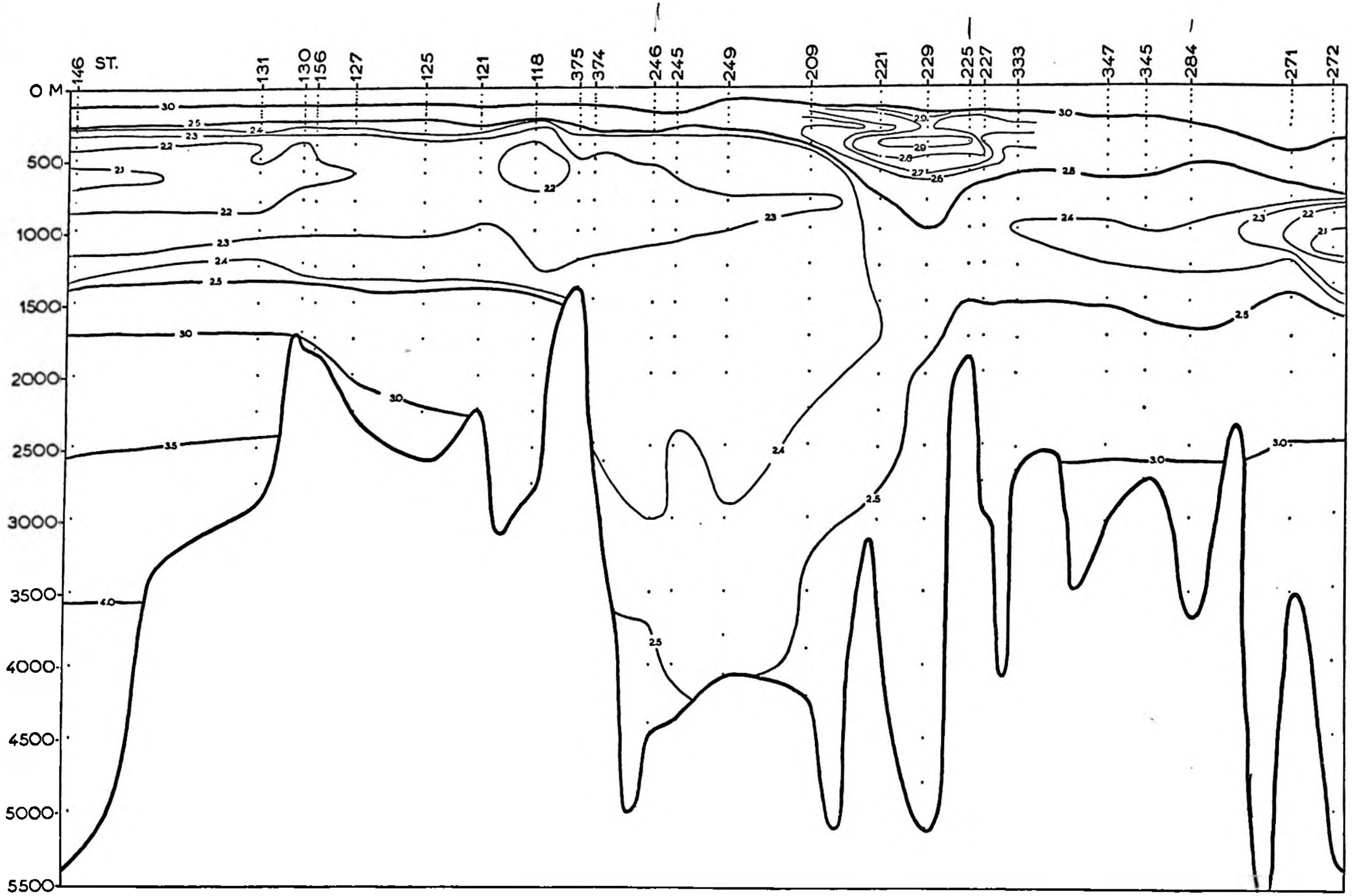

Fig. 18 B. Vertical distribution of oxygen; section II; compare fig. 2.

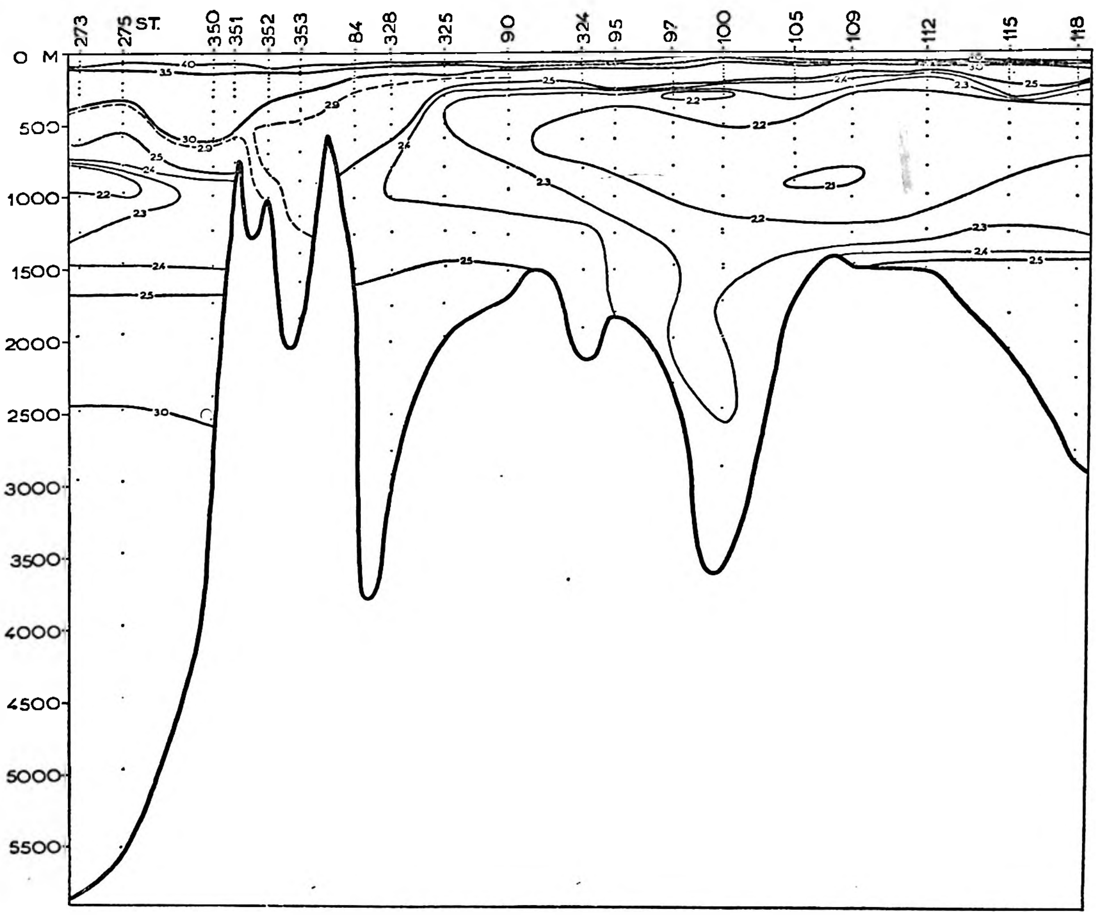

Fig. 18 C. Vertical distribution of oxygen; section III; compare fig. 2.

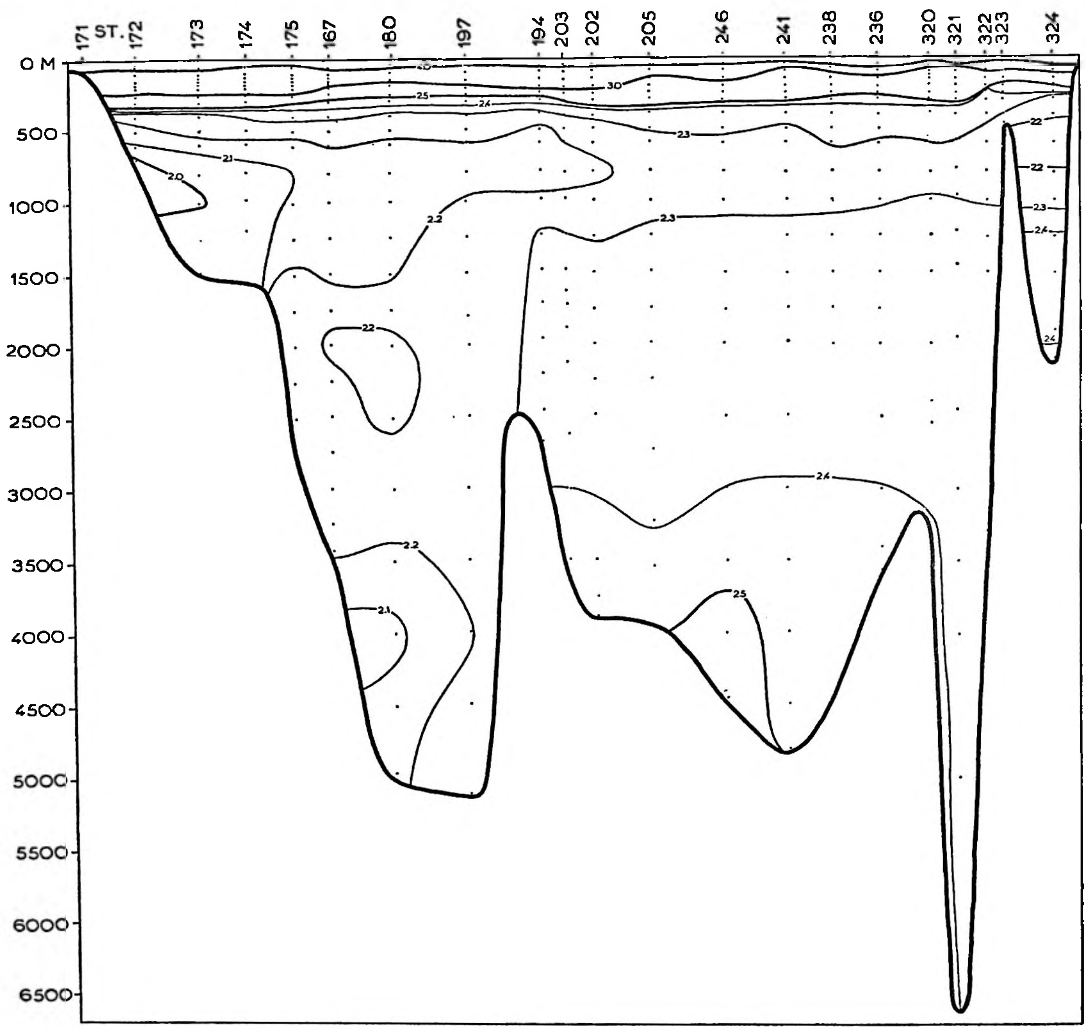

Fig. 18 D. Vertical distribution of oxygen; section IV; compare fig. 2.

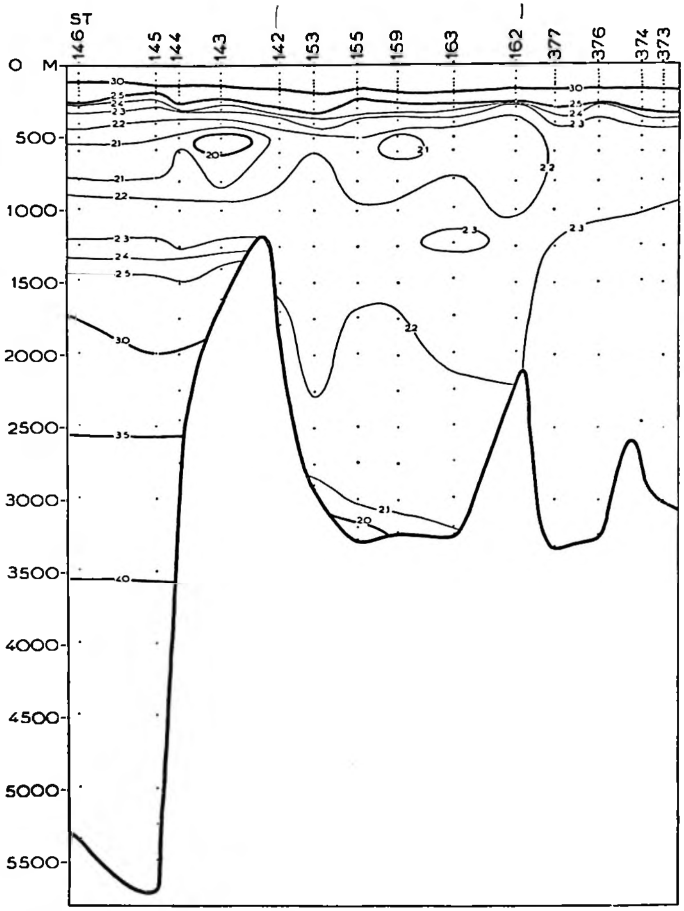

Fig. 18 E. Vertical distribution of oxygen; section V; compare fig. 2.

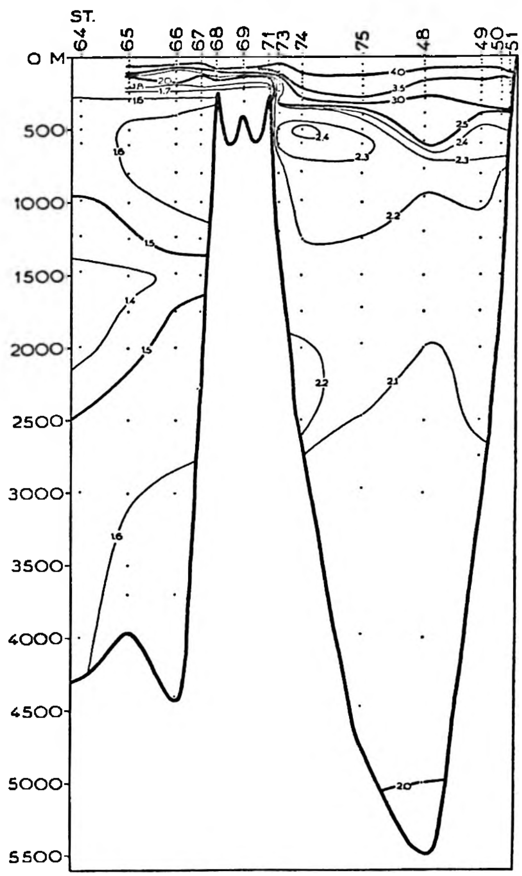

Fig. 18 F. Vertical distribution of oxygen; section VI; compare fig. 2.

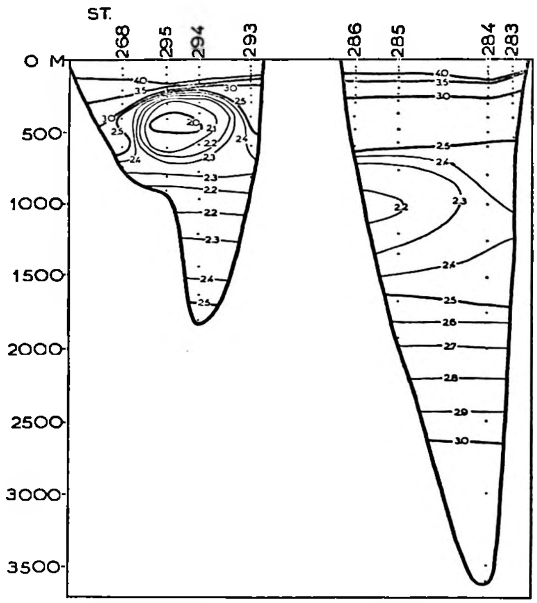

Fig. 18 G. Vertical distribution of oxygen; section VII; compare fig. 2.

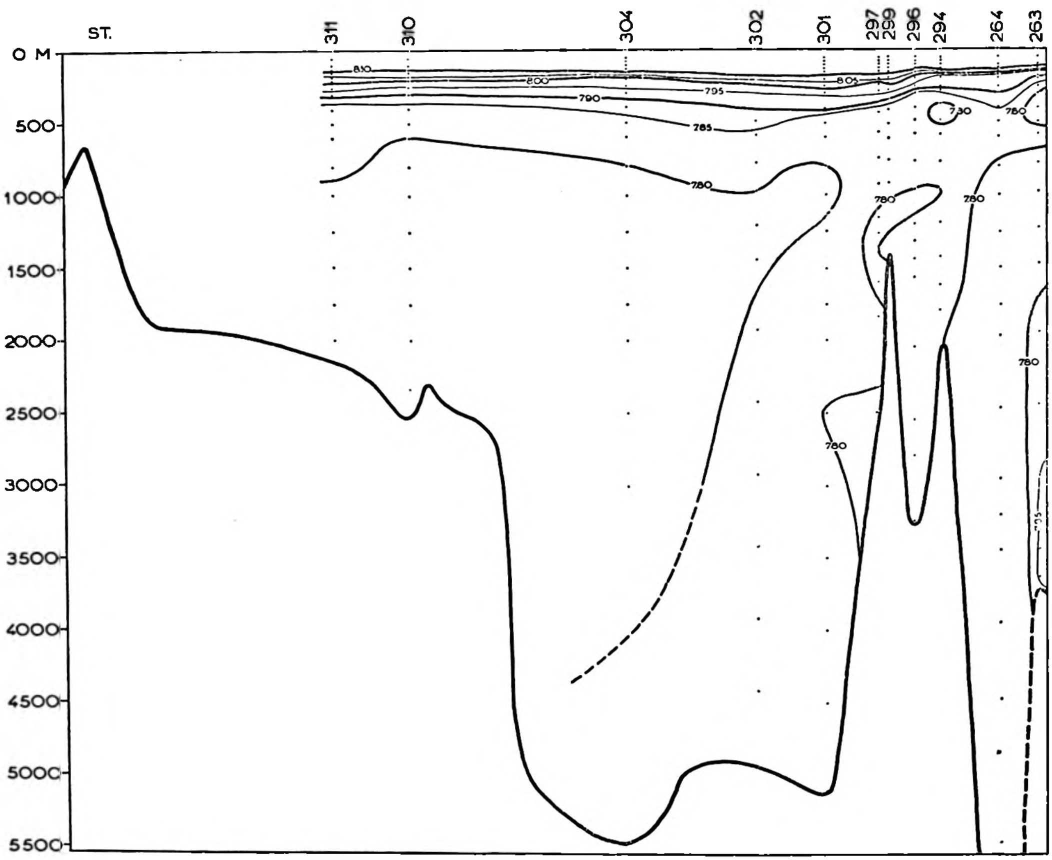

Fig. 19 A. Vertical distribution of pH; section I; compare fig. 2.

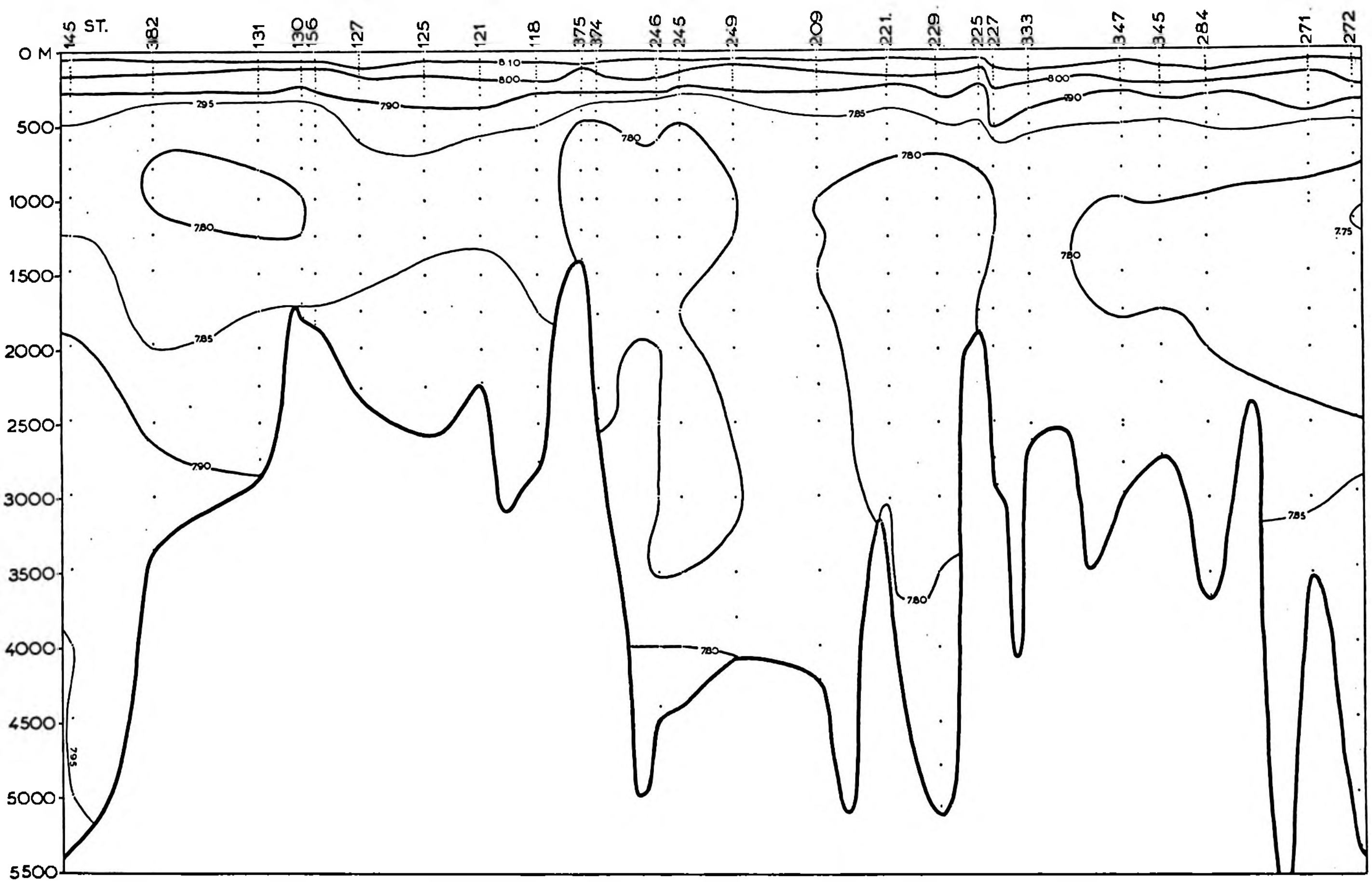

Fig. 19 B. Vertical distribution of pH; section II; compare fig. 2.

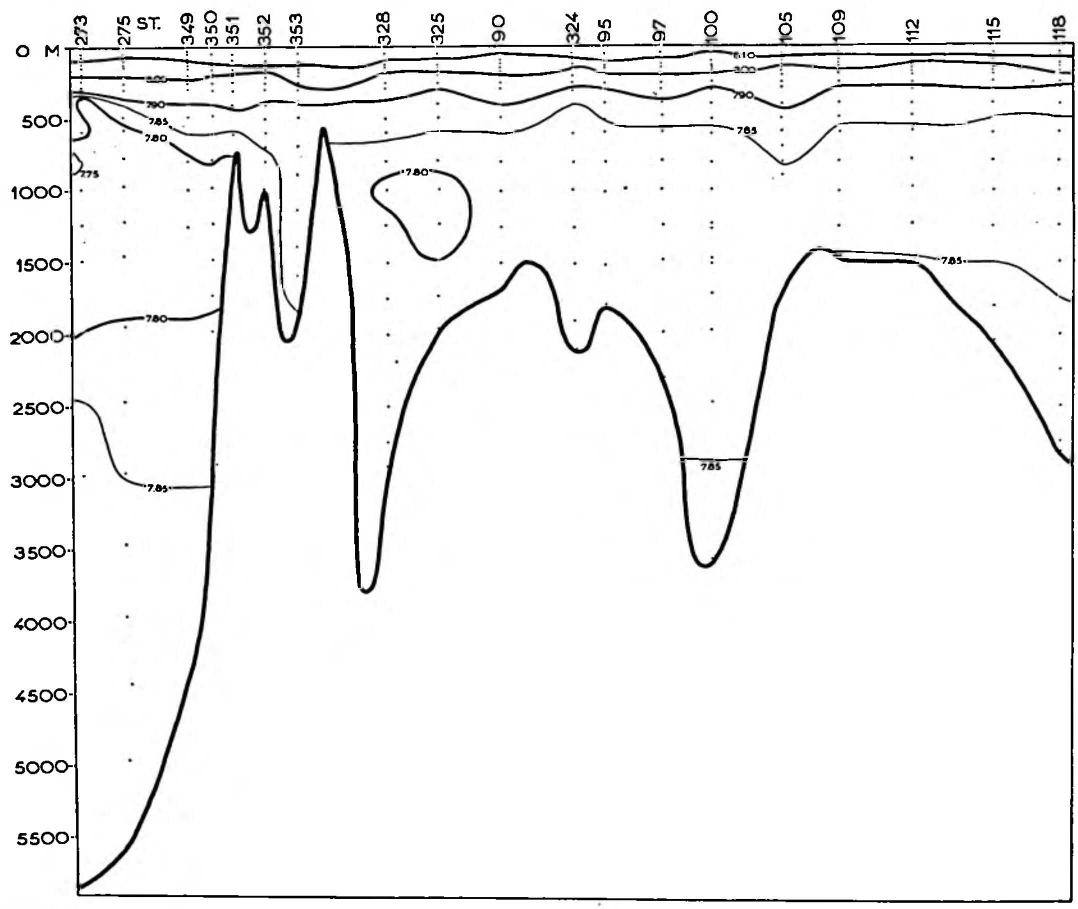

Fig. 19 C. Vertical distribution of pH; section III; compare fig. 2.

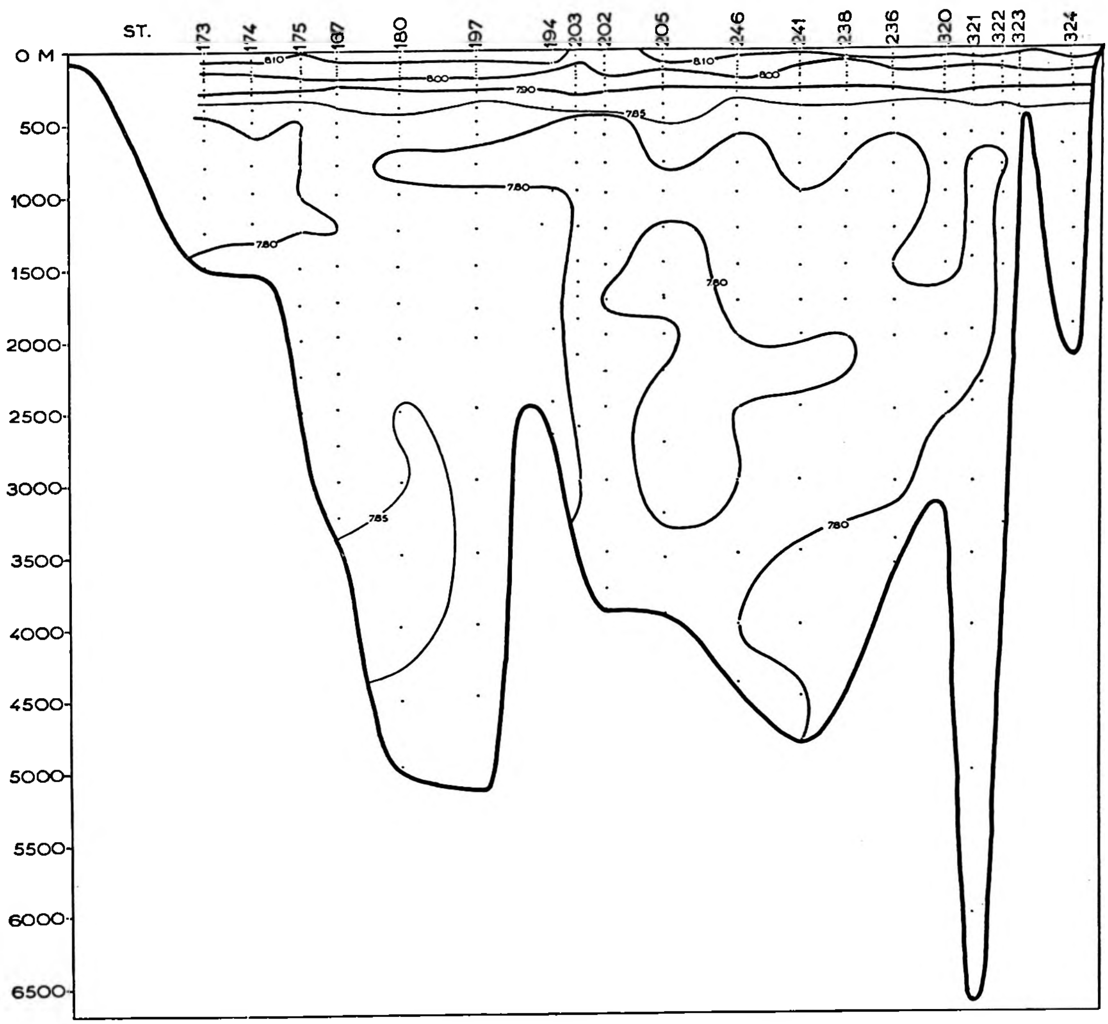

Fig. 19 D. Vertical distribution of pH; section IV; compare fig. 2.

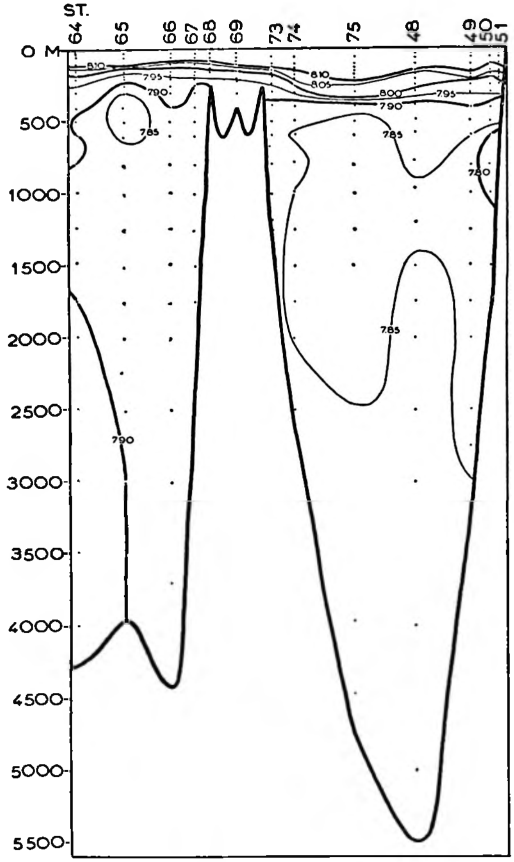

Fig. 19 E. Vertical distribution of pH; section VI; compare fig. 2.

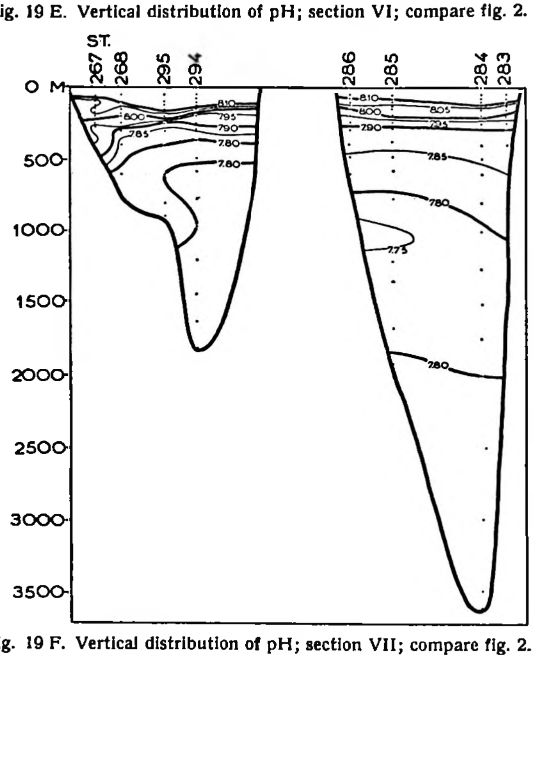

Fig. 19 F. Vertical distribution of pH; section VII; compare fig. 2.

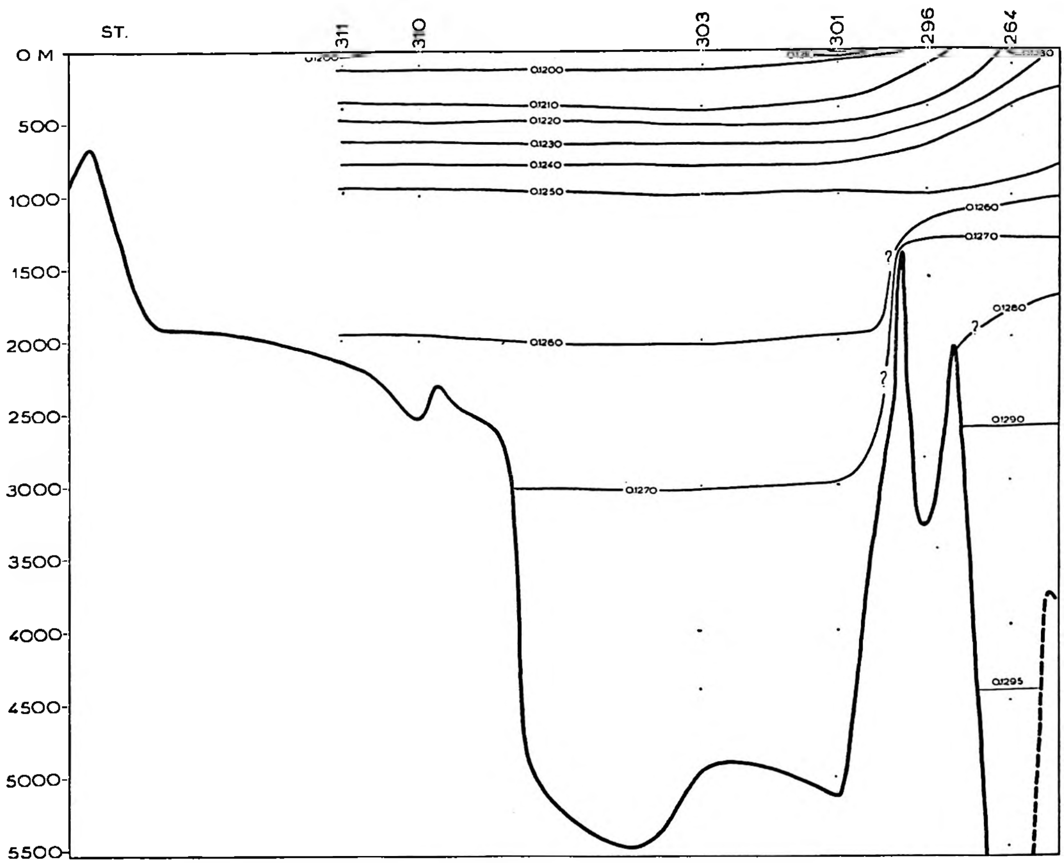

Fig. 20 A. Vertical distribution of specific alkalinity; section I, compare fig. 2.

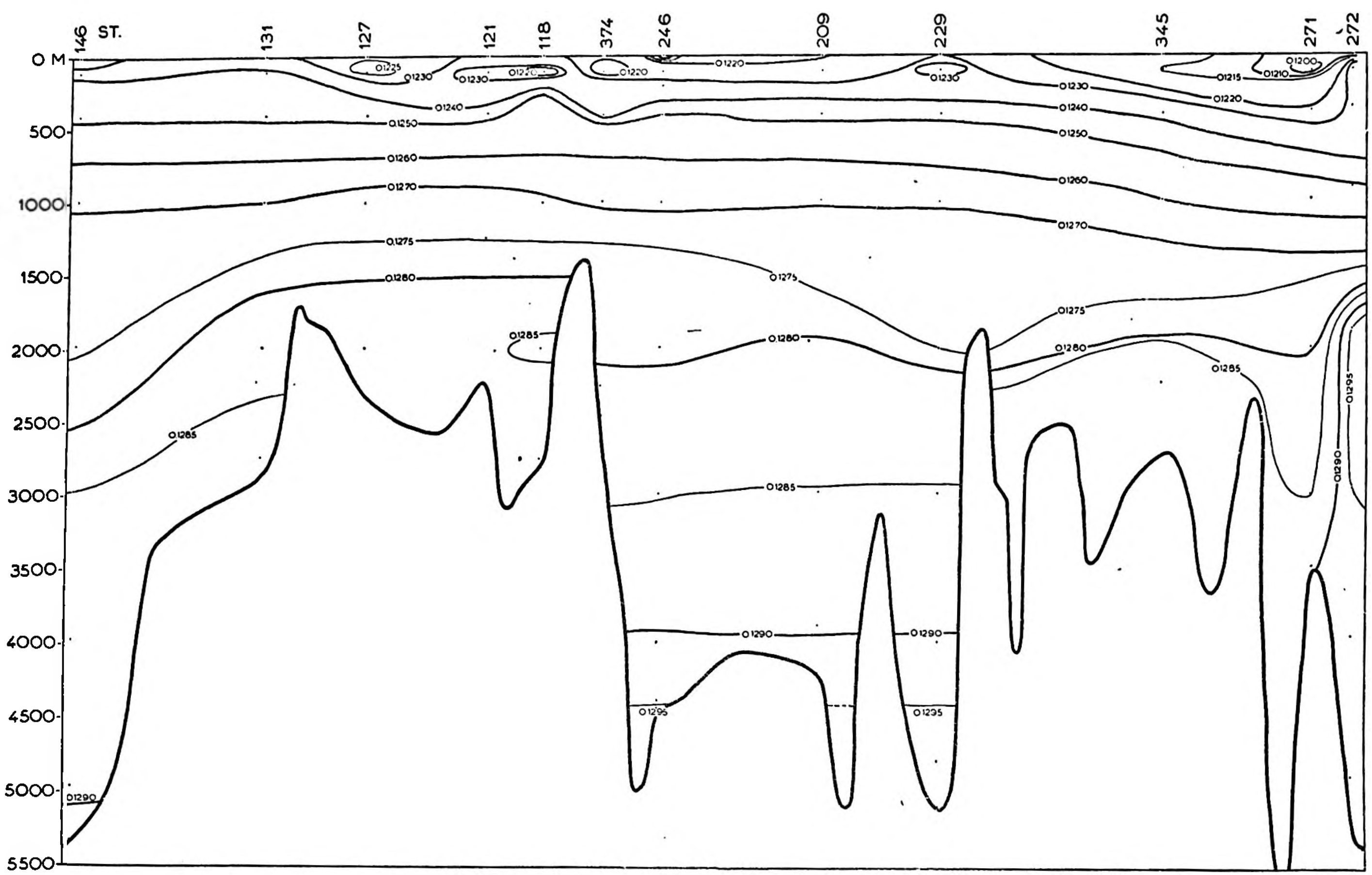

Fig. 20 B. Vertical distribution of specific alkalinity; section II; compare fig. 2.

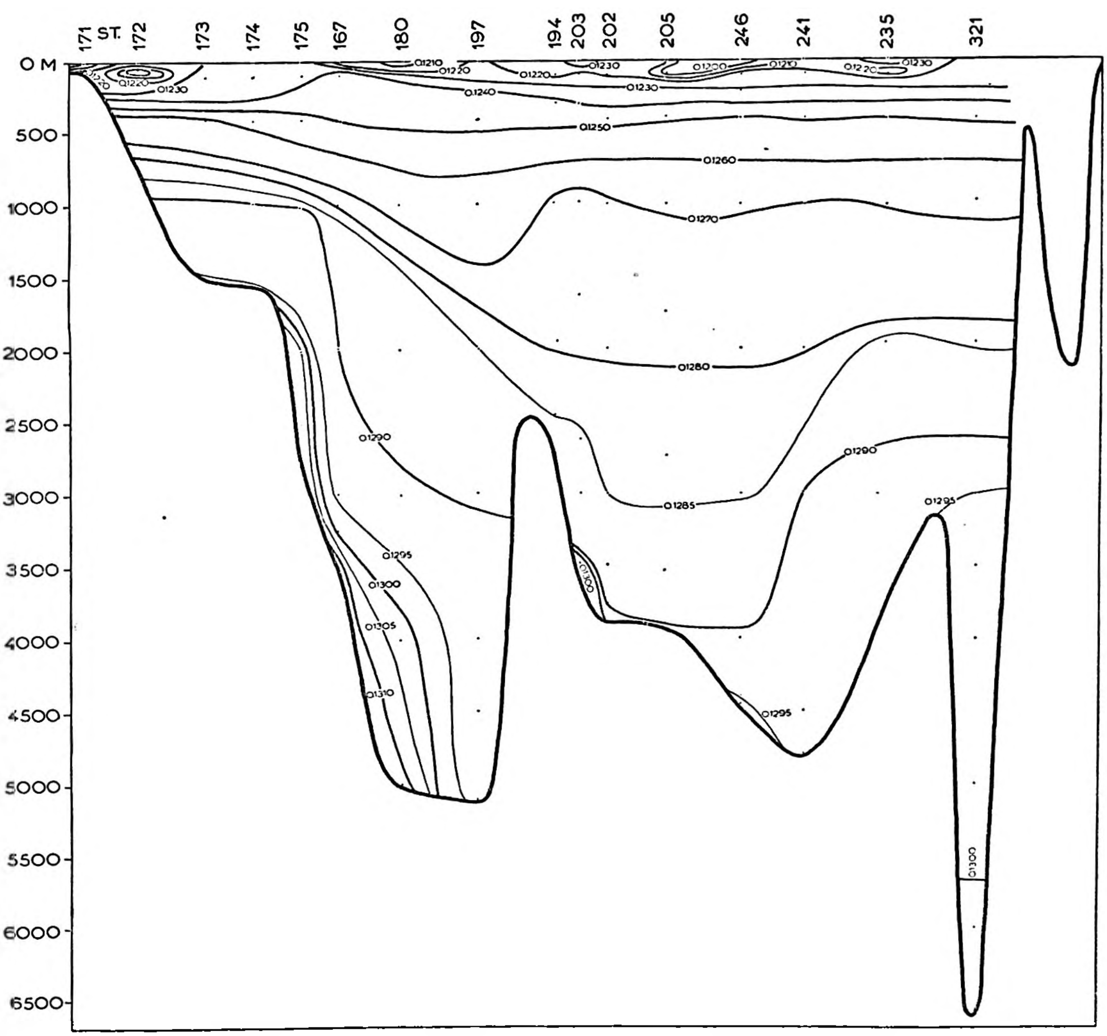

Fig. 20 C. Vertical distribution of specific alkalinity; section IV; compare fig. 2.

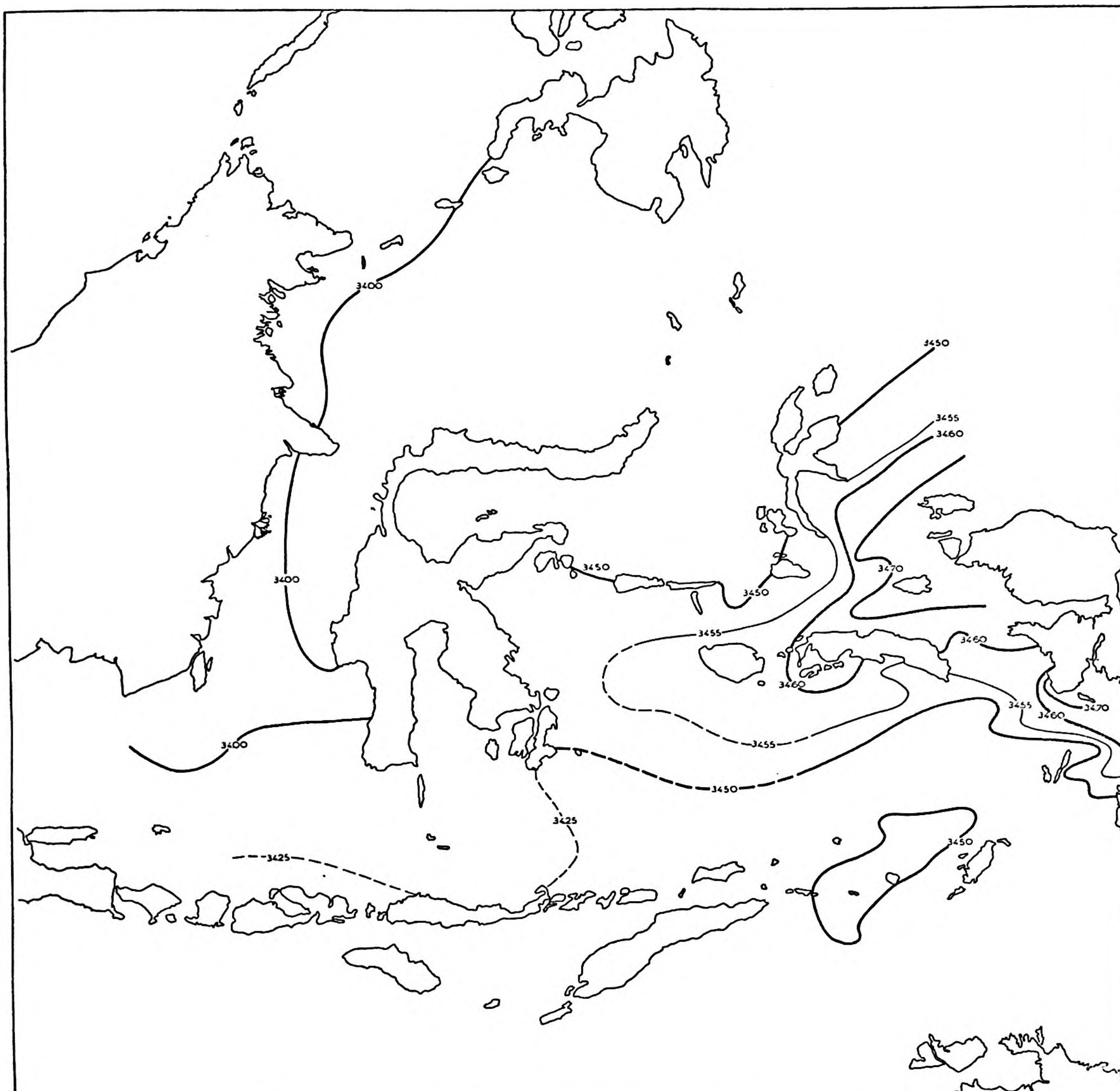

Fig. 21 A. Horizontal distribution of salinity; 50 m.

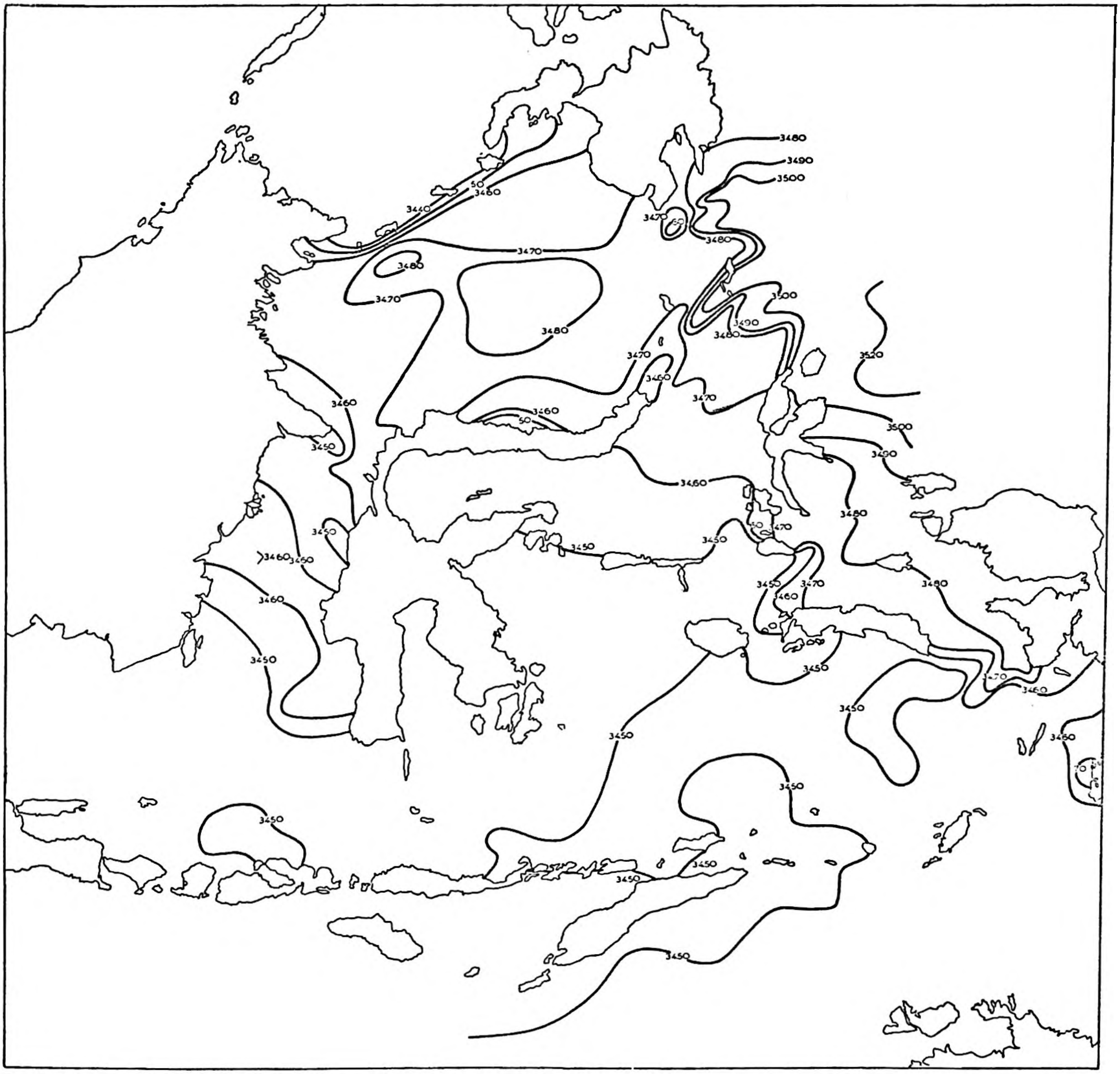

Fig. 21 B. Horizontal distribution of salinity; 100 m.

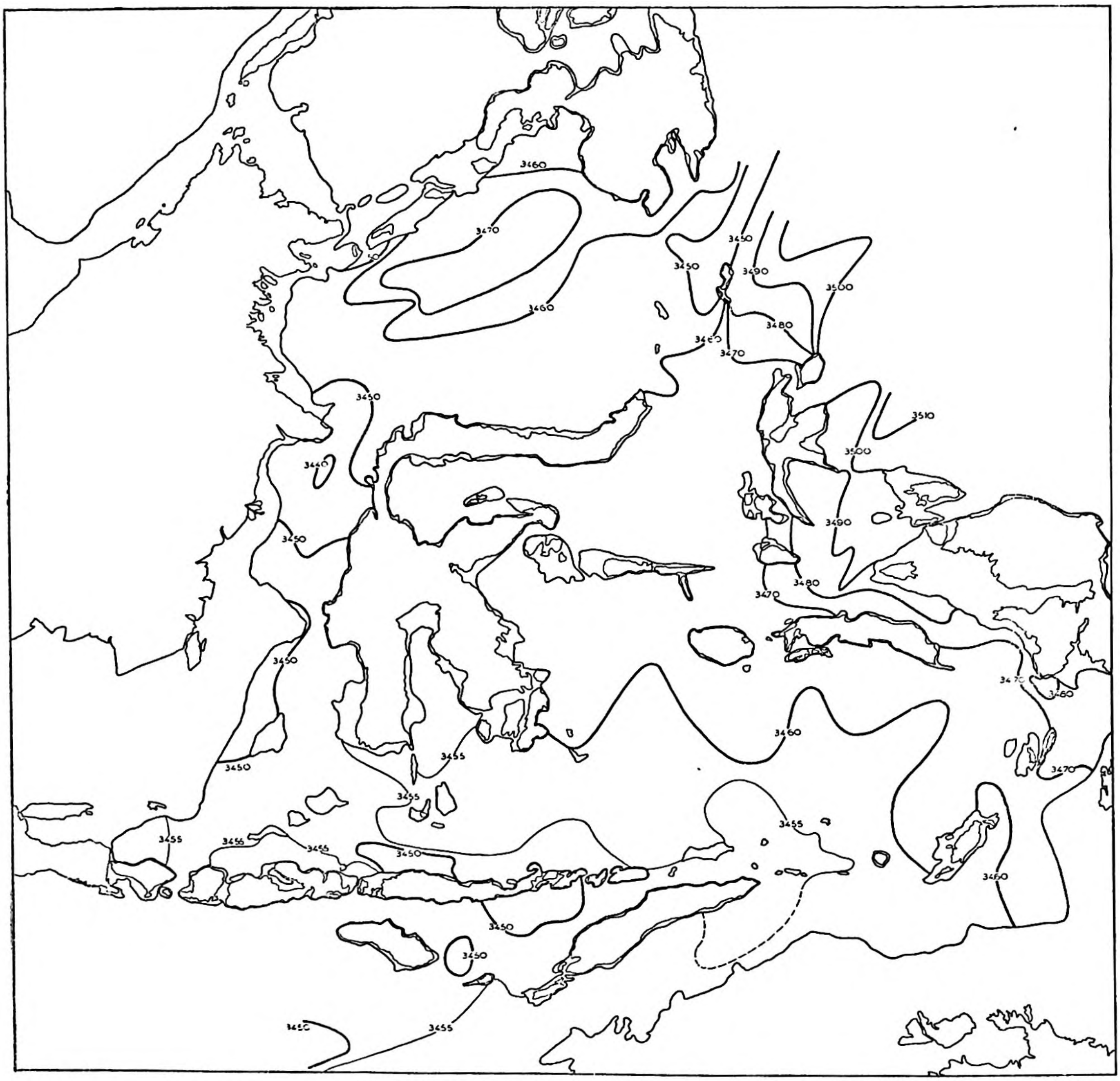

Fig. 21 C. Horizontal distribution of salinity; 200 m.

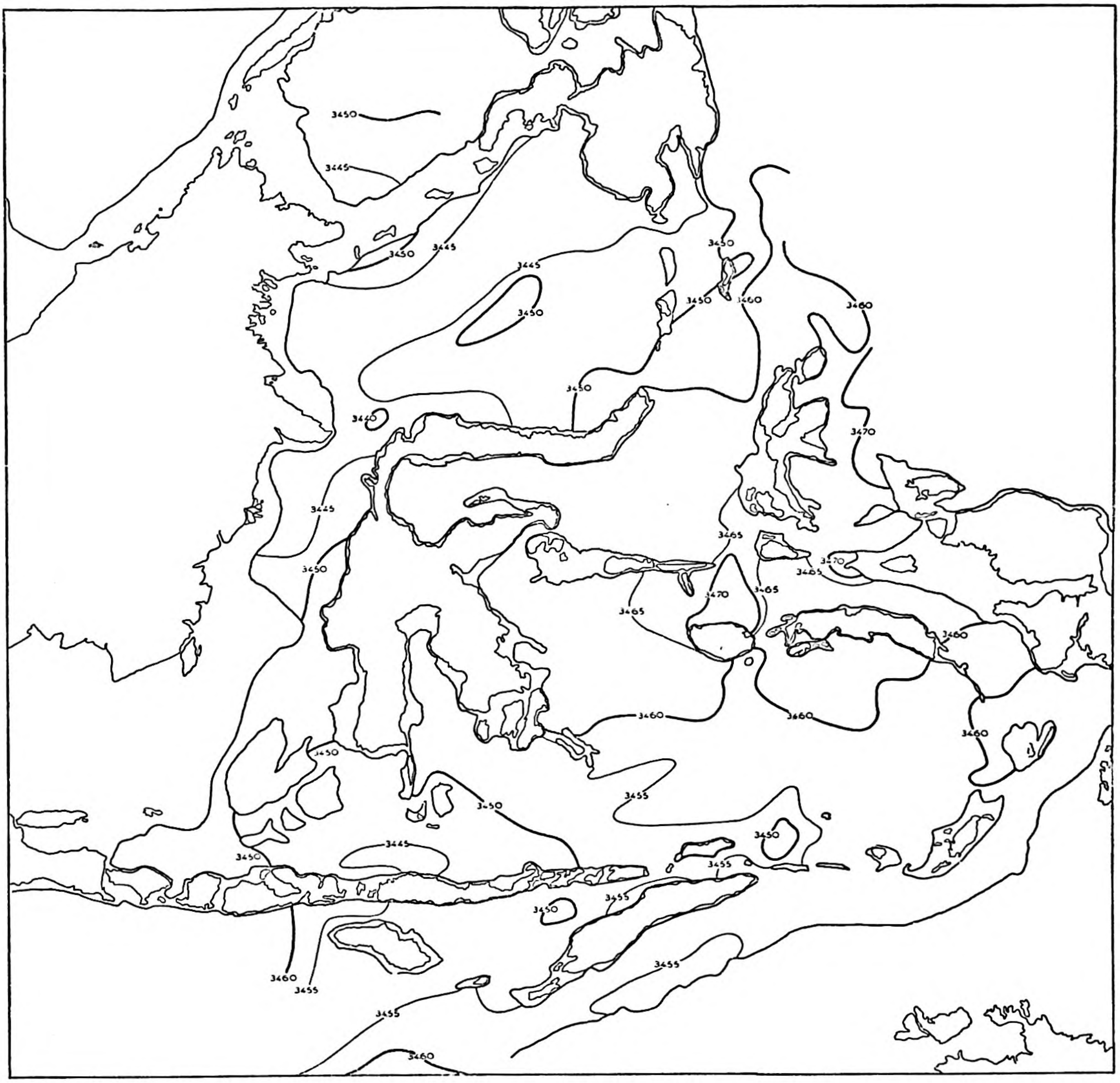

Fig. 21 D. Horizontal distribution of salinity; 400 m.

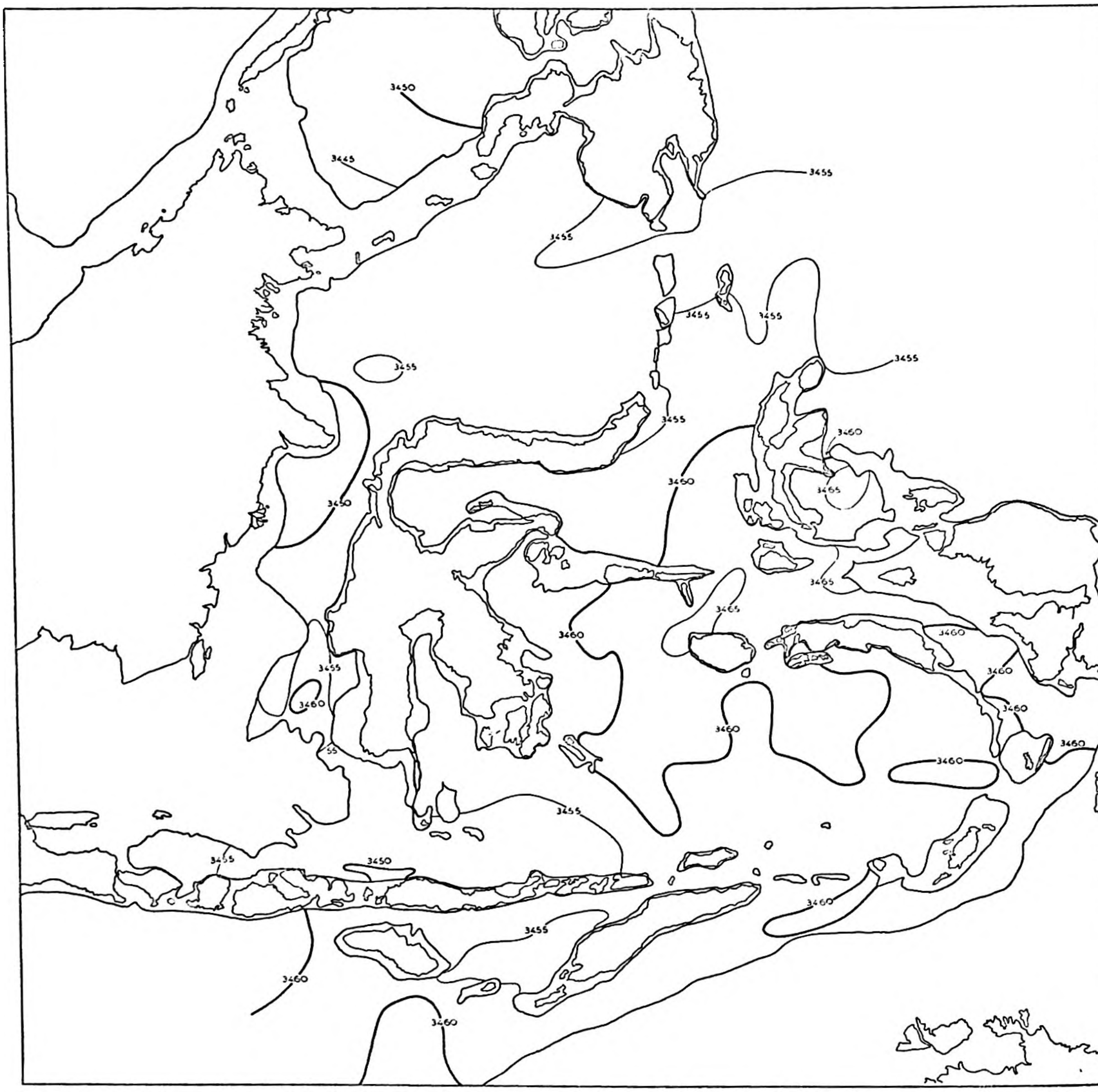

Fig. 21 E. Horizontal distribution of salinity; 600 m.

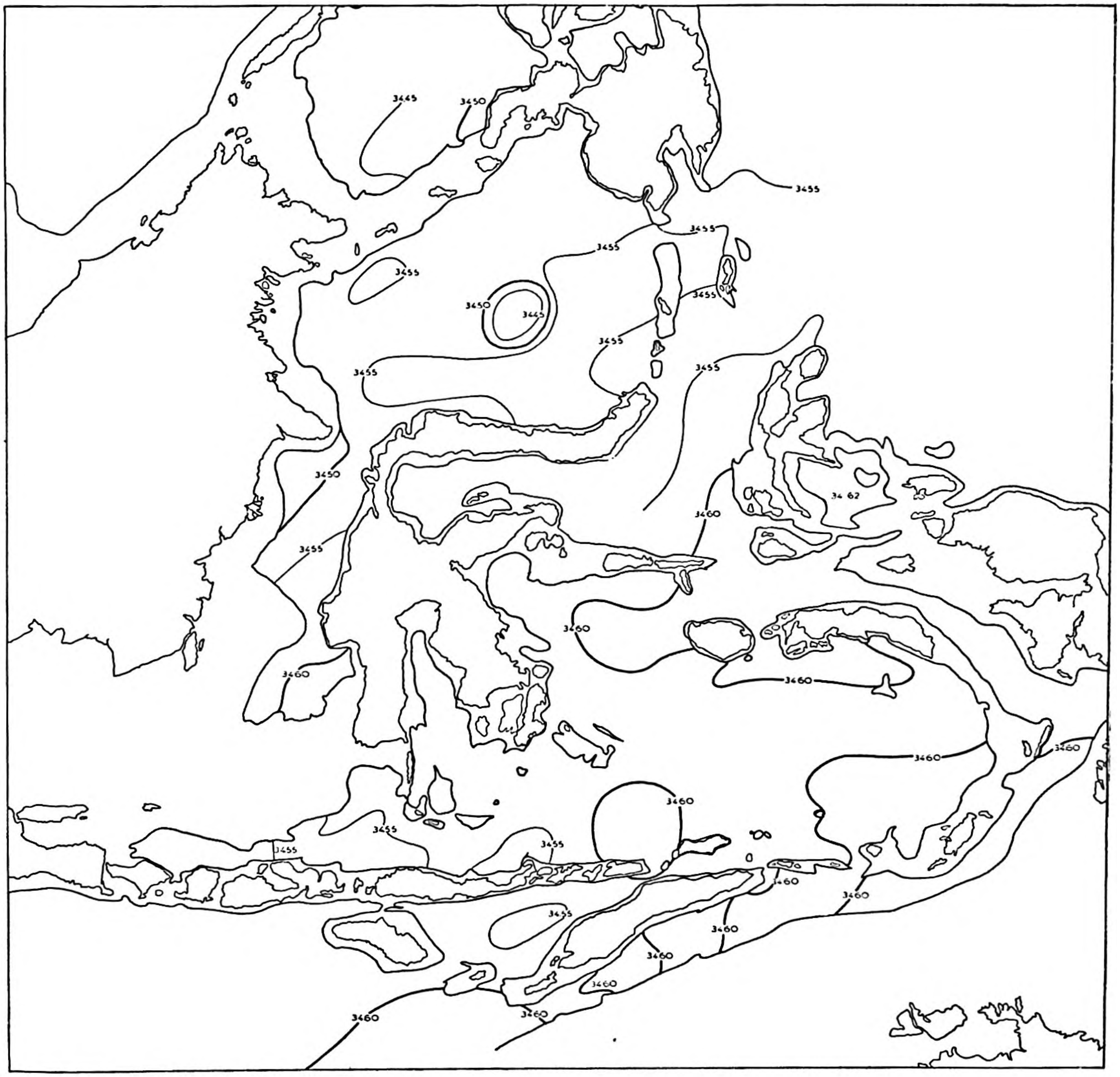

Fig. 21 F. Horizontal distribution of salinity; 800 m.

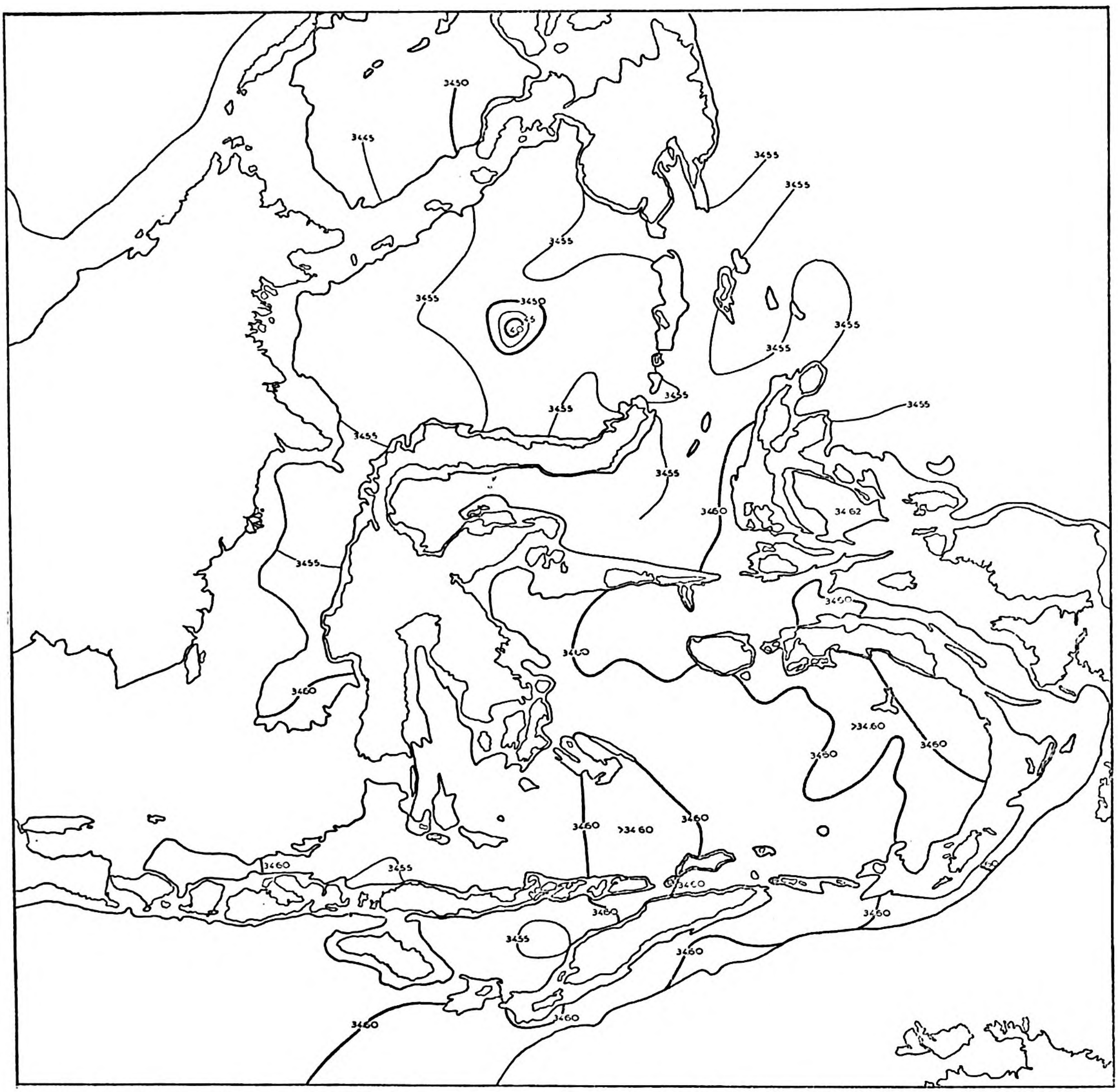

Fig. 21 G. Horizontal distribution of salinity; 1000 m.

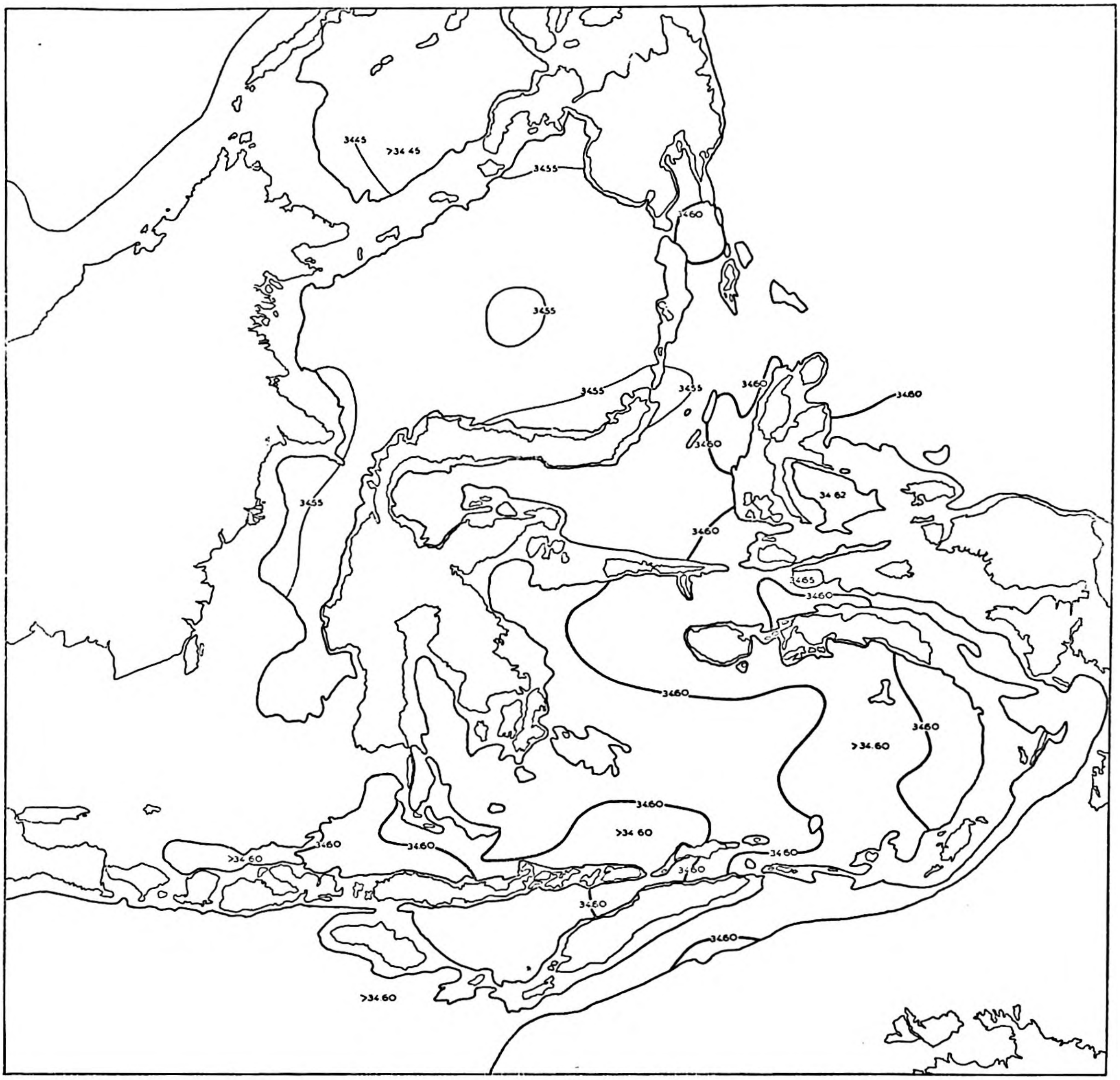

Fig. 21 H. Horizontal distribution of salinity; 1250 m.

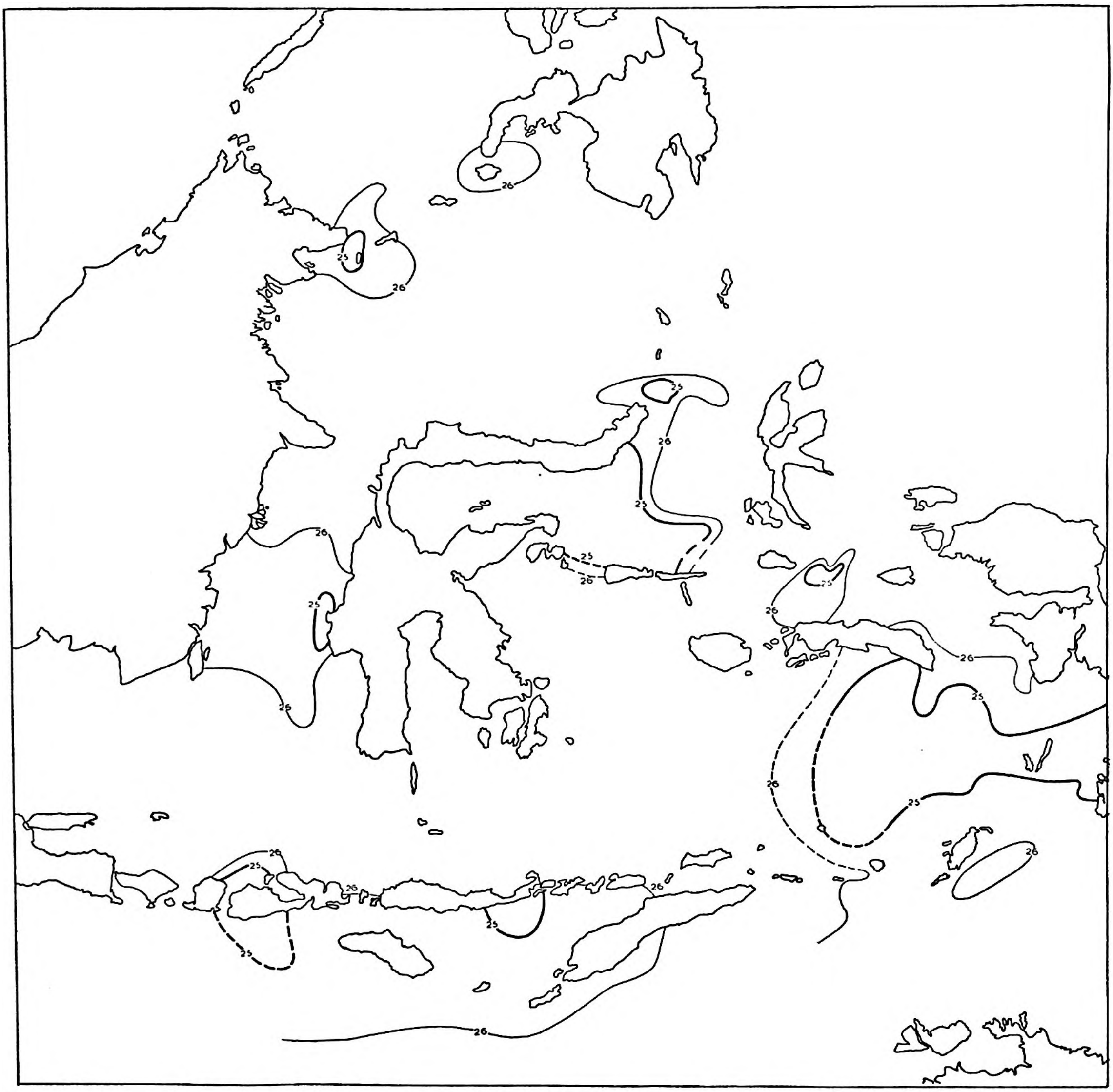

Fig. 22 A. Horizontal distribution of temperature; 50 m.

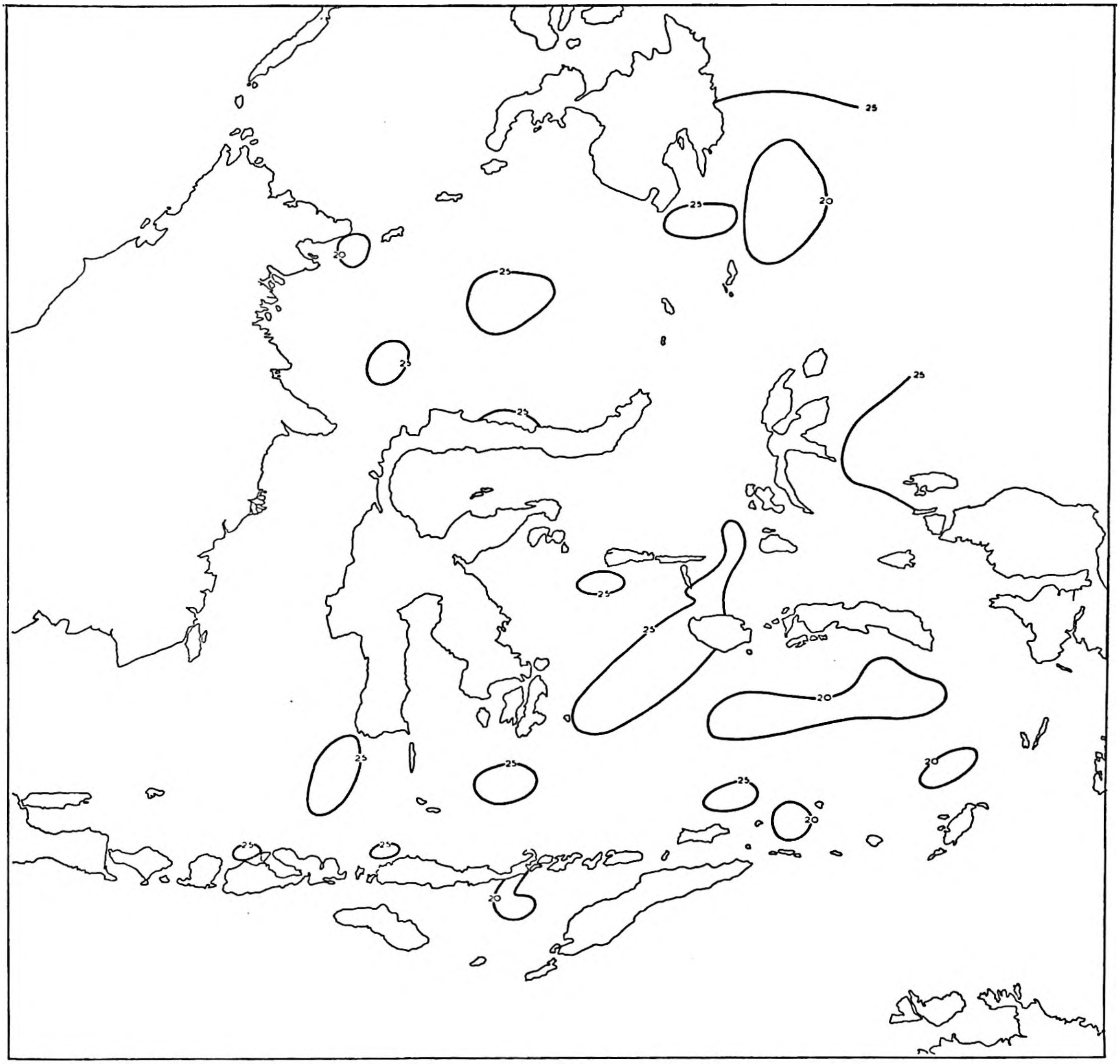

Fig. 22 B. Horizontal distribution of temperature; 100 m.

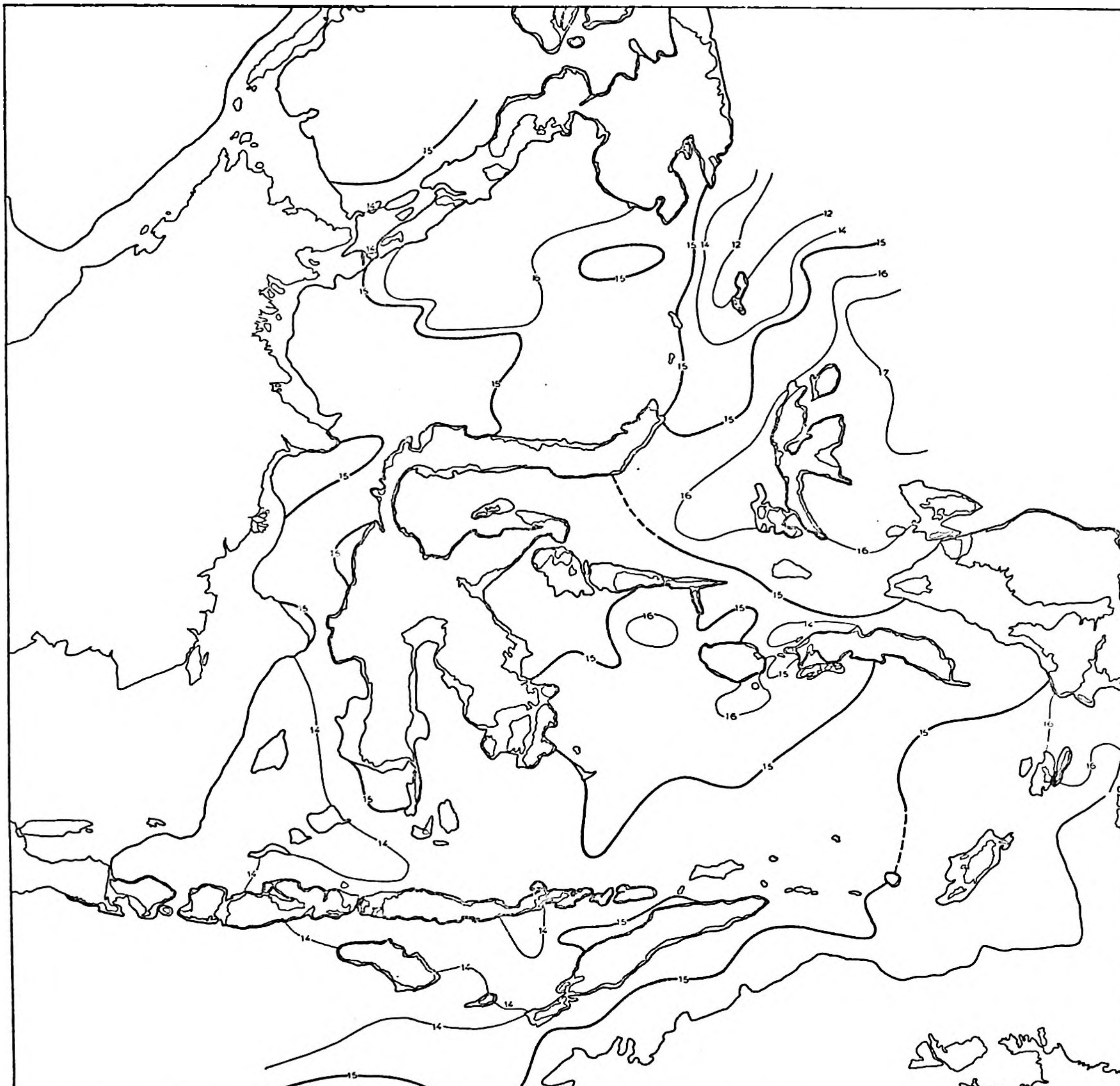

Fig. 22 C. Horizontal distribution of temperature; 200 m.

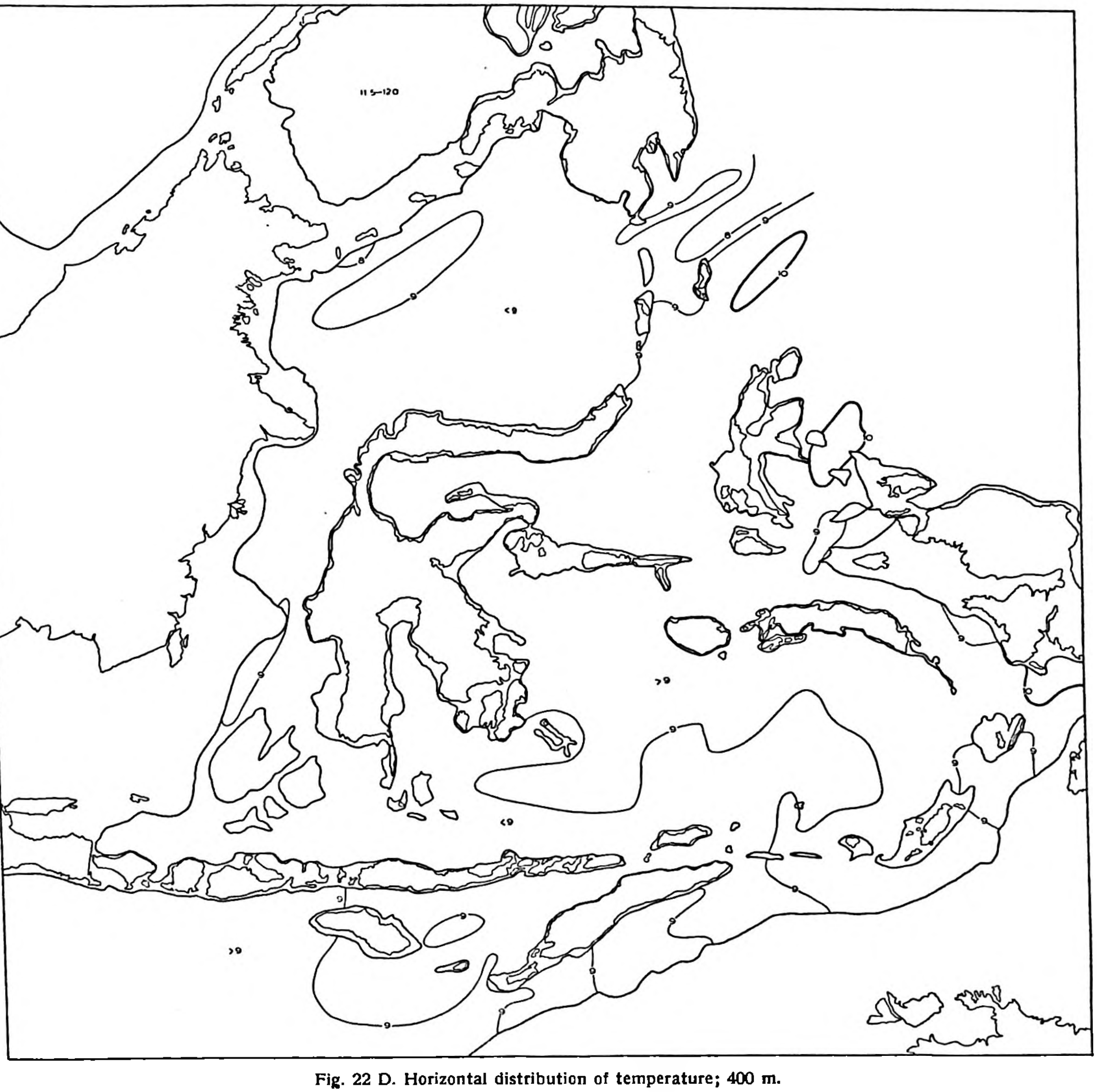

Fig. 22 D. Horizontal distribution of temperature; 400 m.

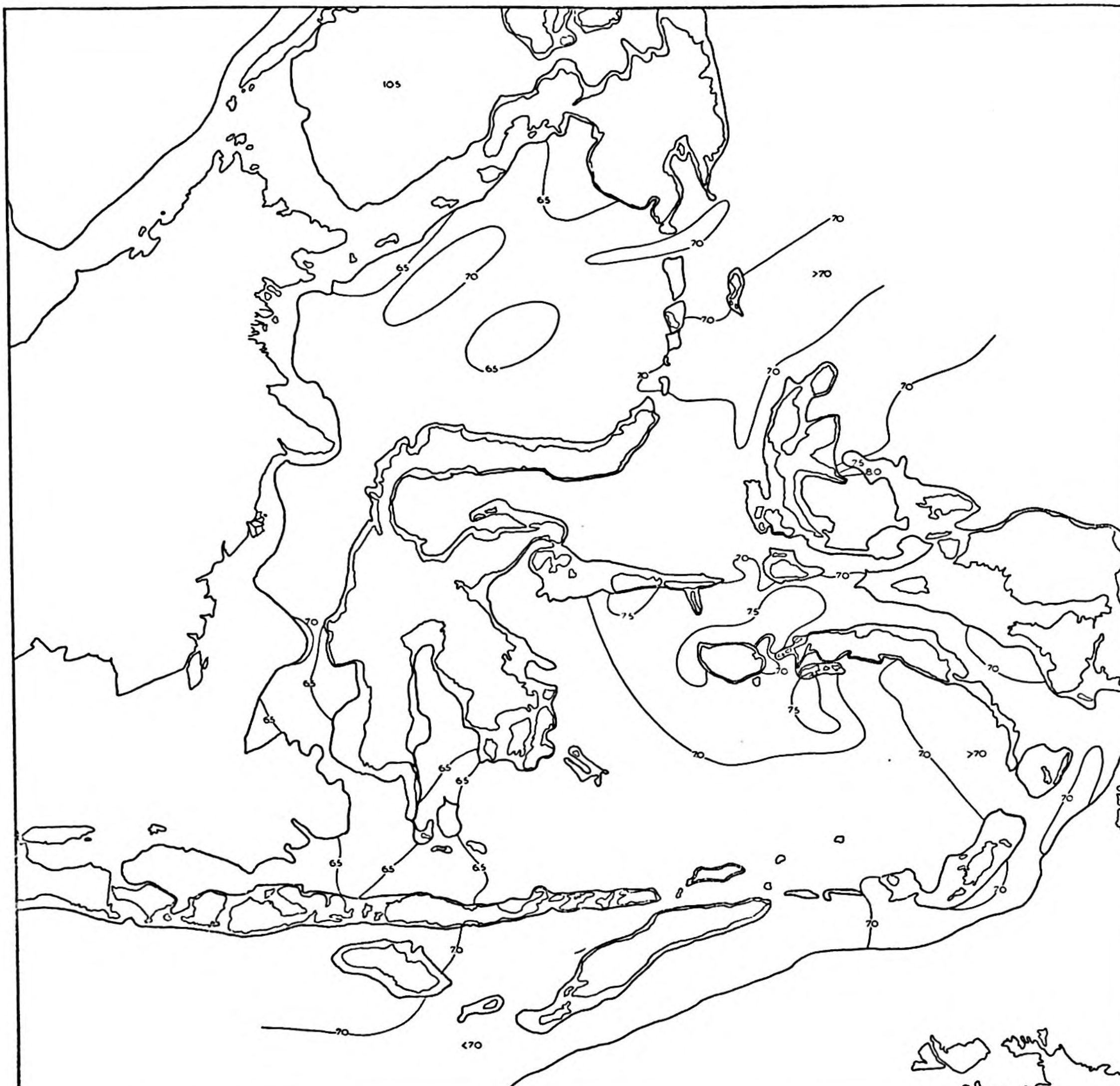

Fig. 22 E. Horizontal distribution of temperature; 600 m.

Fig. 22 F. Horizontal distribution of temperature; 800 m.

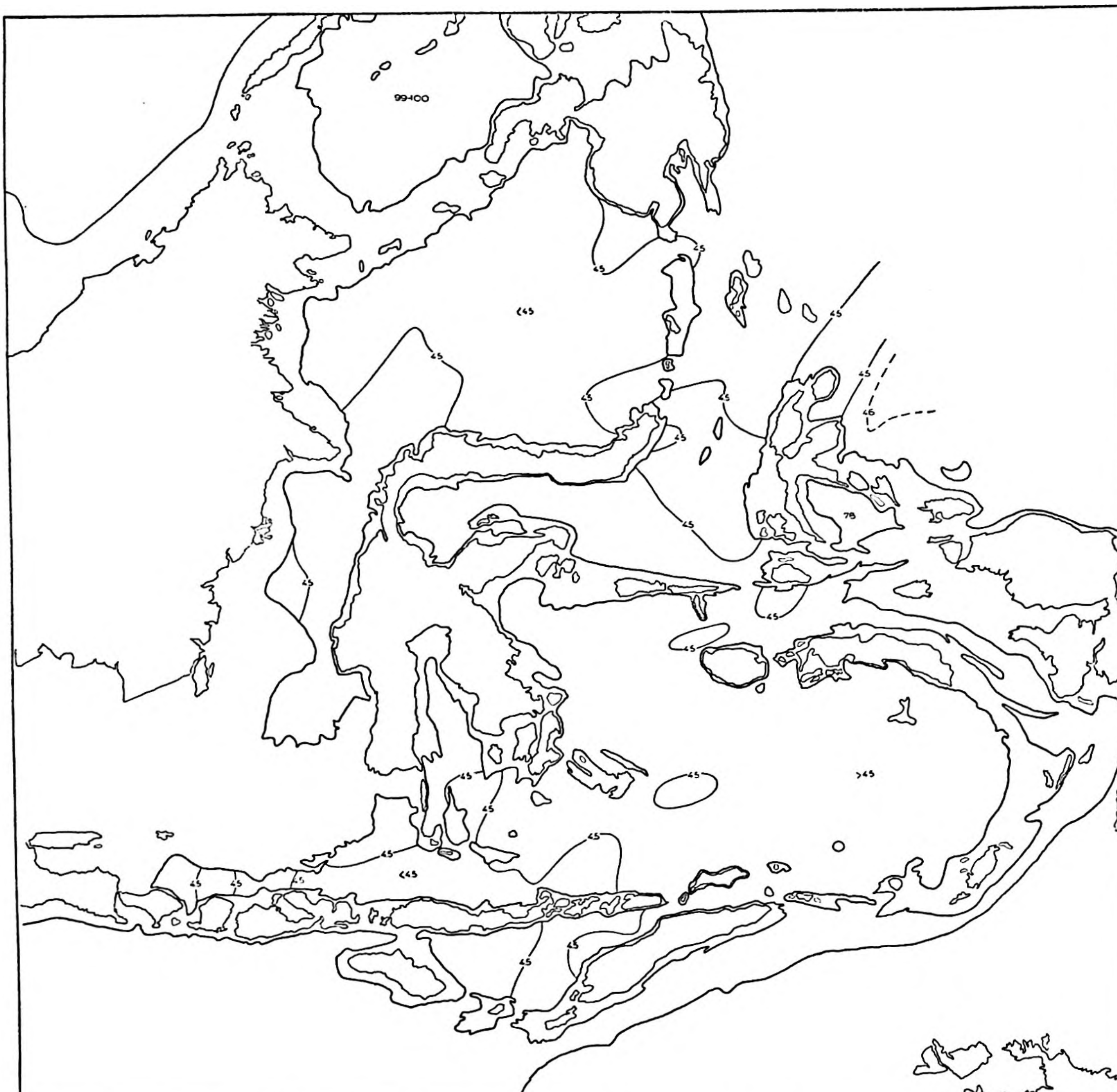

Fig. 22 G. Horizontal distribution of temperature; 1000 m.

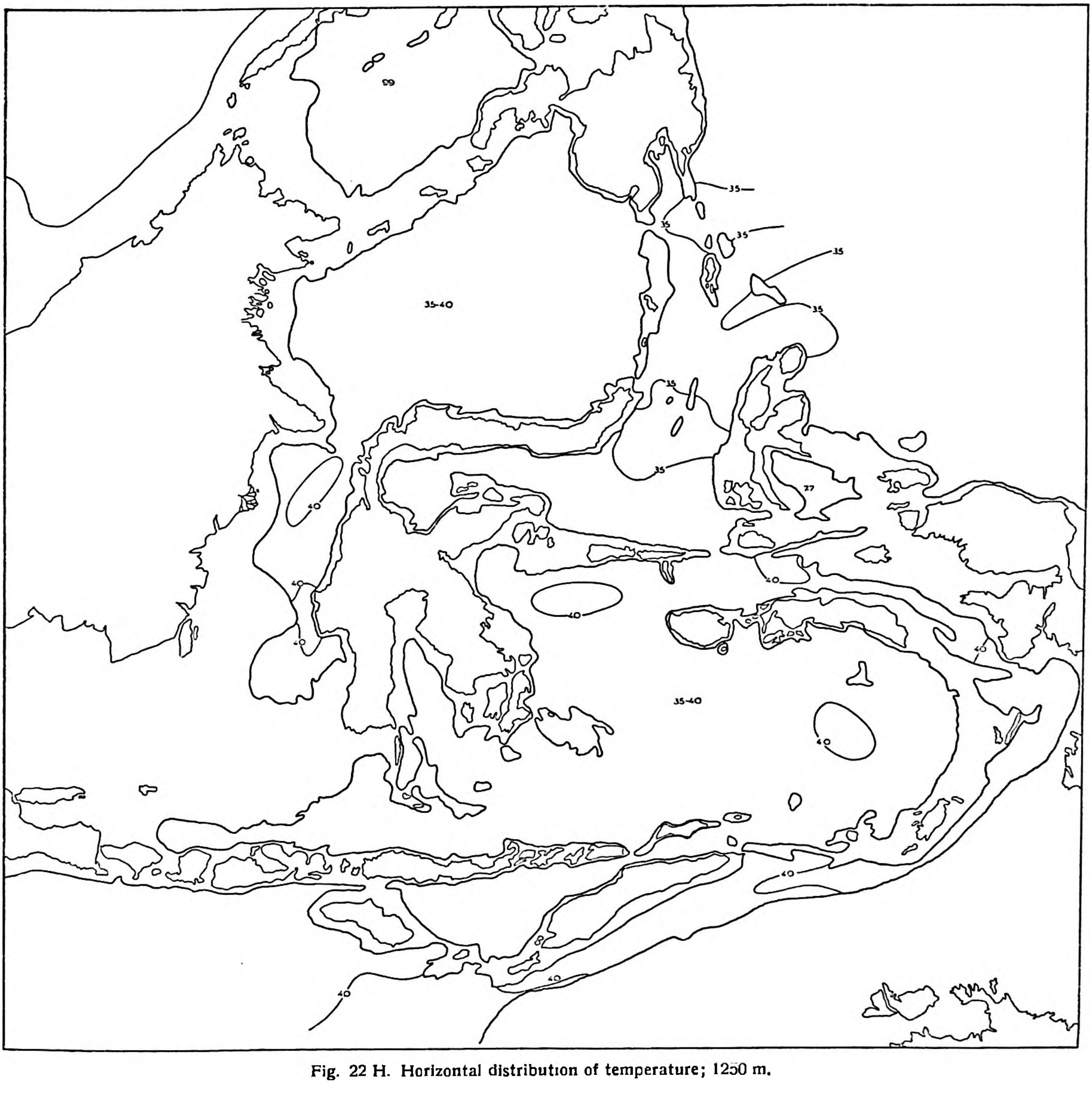

Fig. 22 H. Horizontal distribution of temperature; 1250 m.

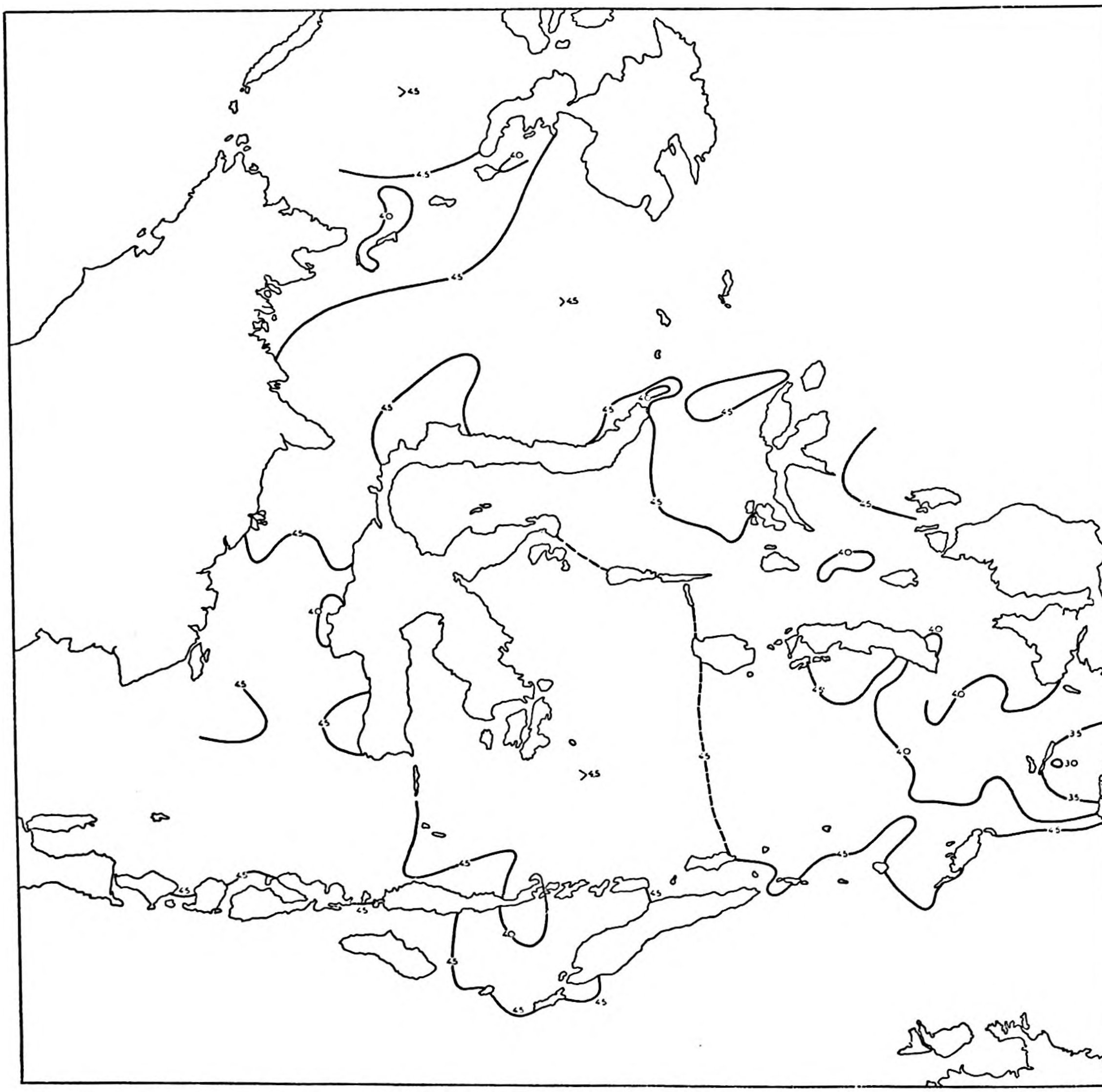

Fig. 23 A. Horizontal distribution of oxygen; 50 m.

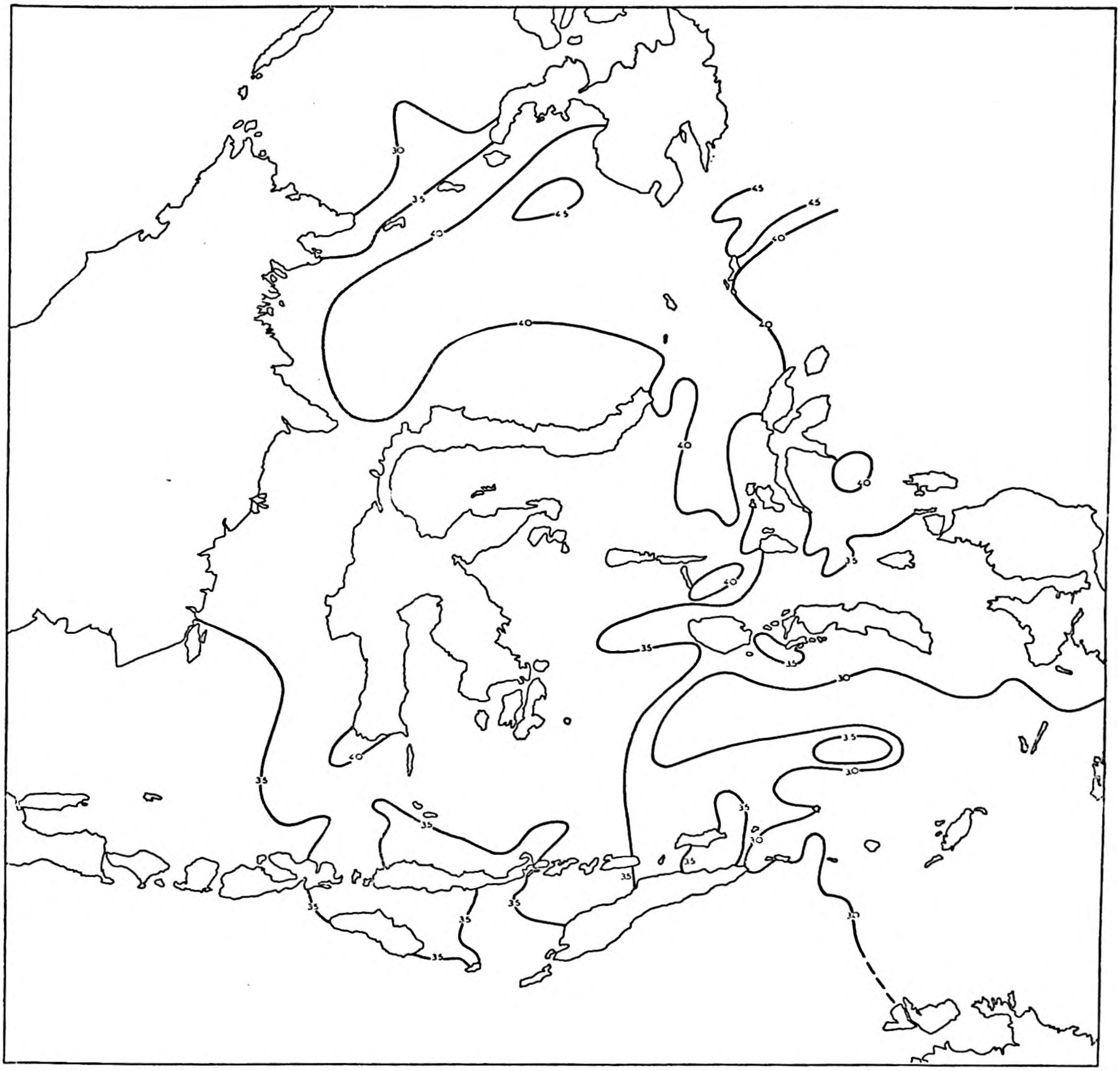

Fig. 23 B. Horizontal distribution of oxygen; 100 m.

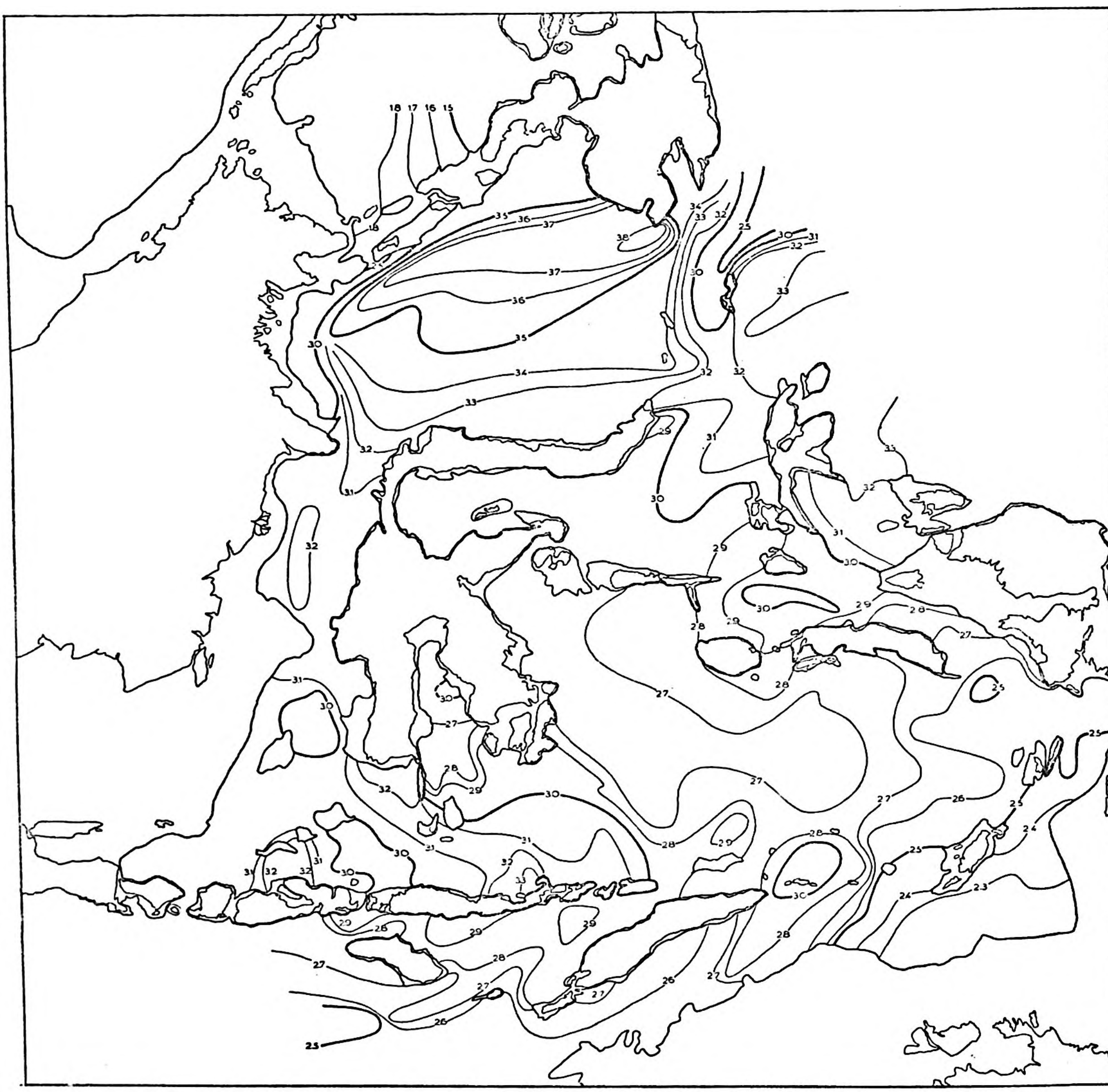

Fig. 23 C. Horizontal distribution of oxygen; 200 m.

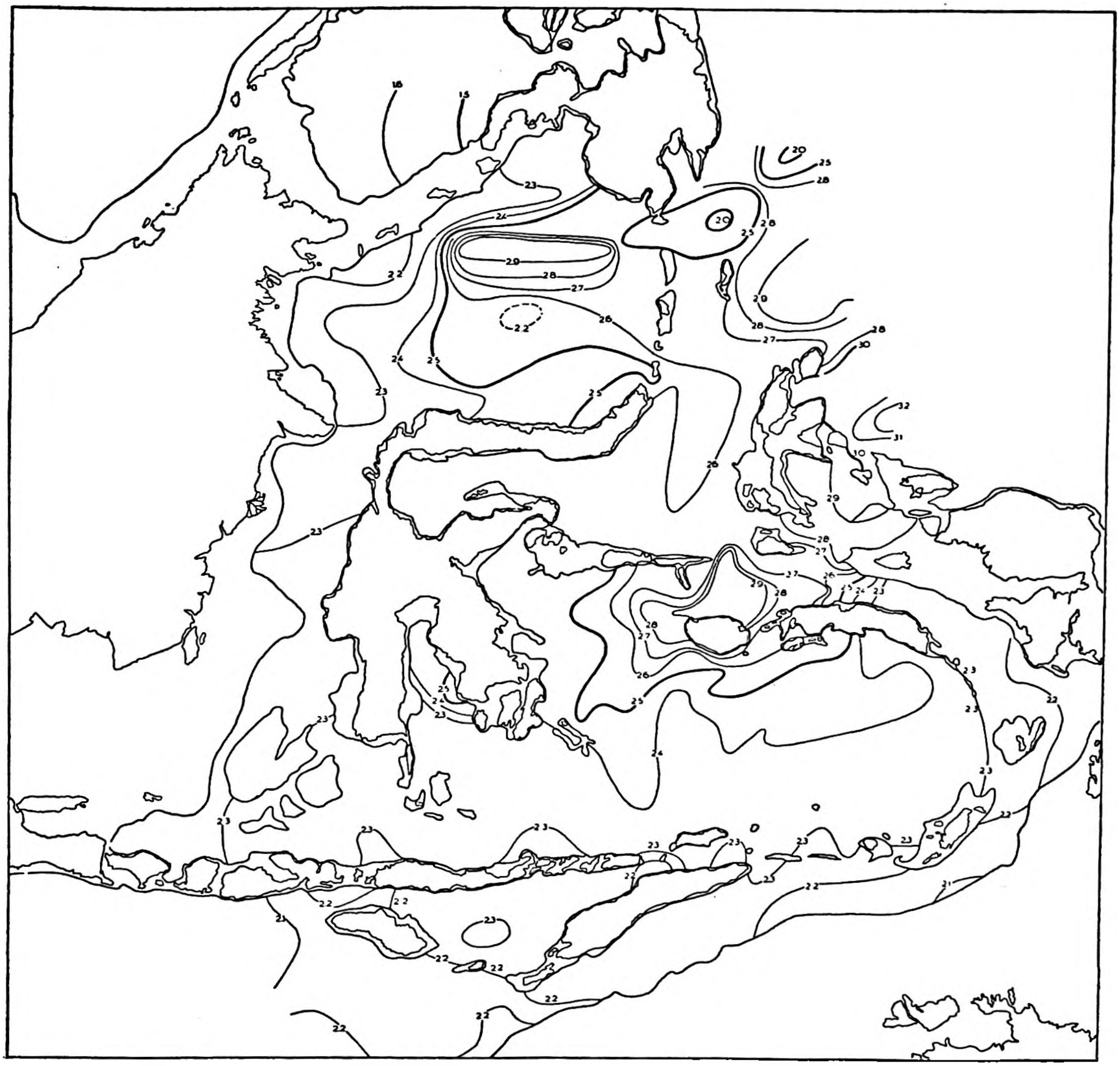

Fig. 23 D. Horizontal distribution of oxygen; 400 m.

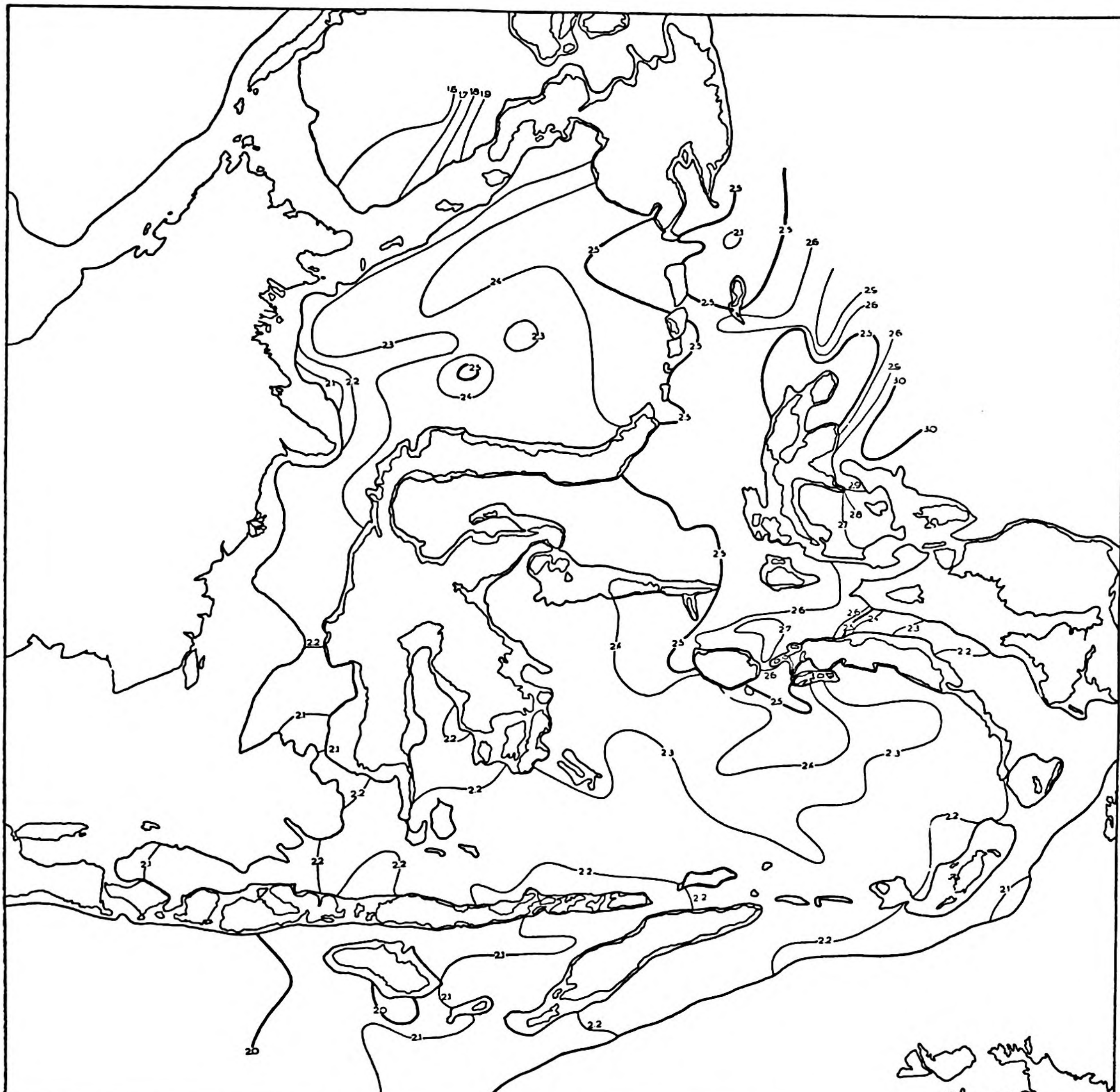

Fig. 23 E. Horizontal distribution of oxygen; 600 m.

Fig. 23 F. Horizontal distribution of oxygen; 800 m.

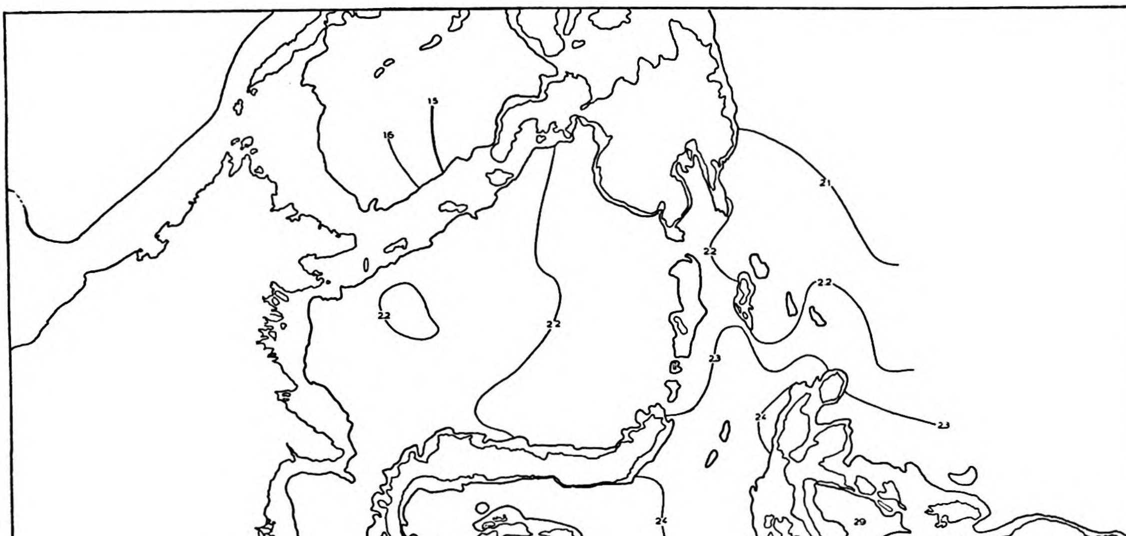

Fig. 23 G. Horizontal distribution of oxygen; 1000 m.

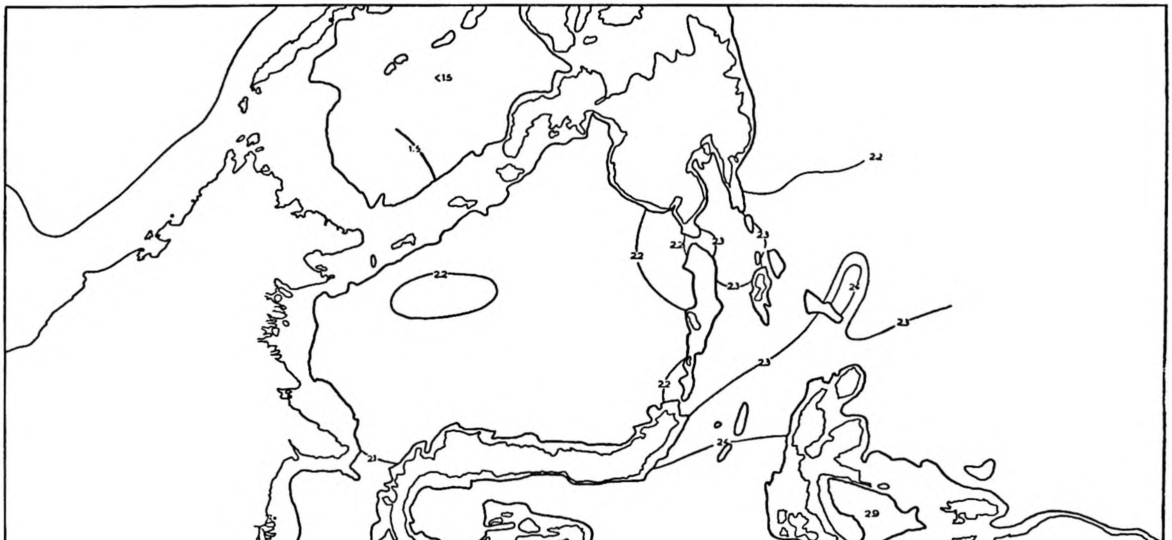

Fig. 23 H. Horizontal distribution of oxygen; 1250 m.

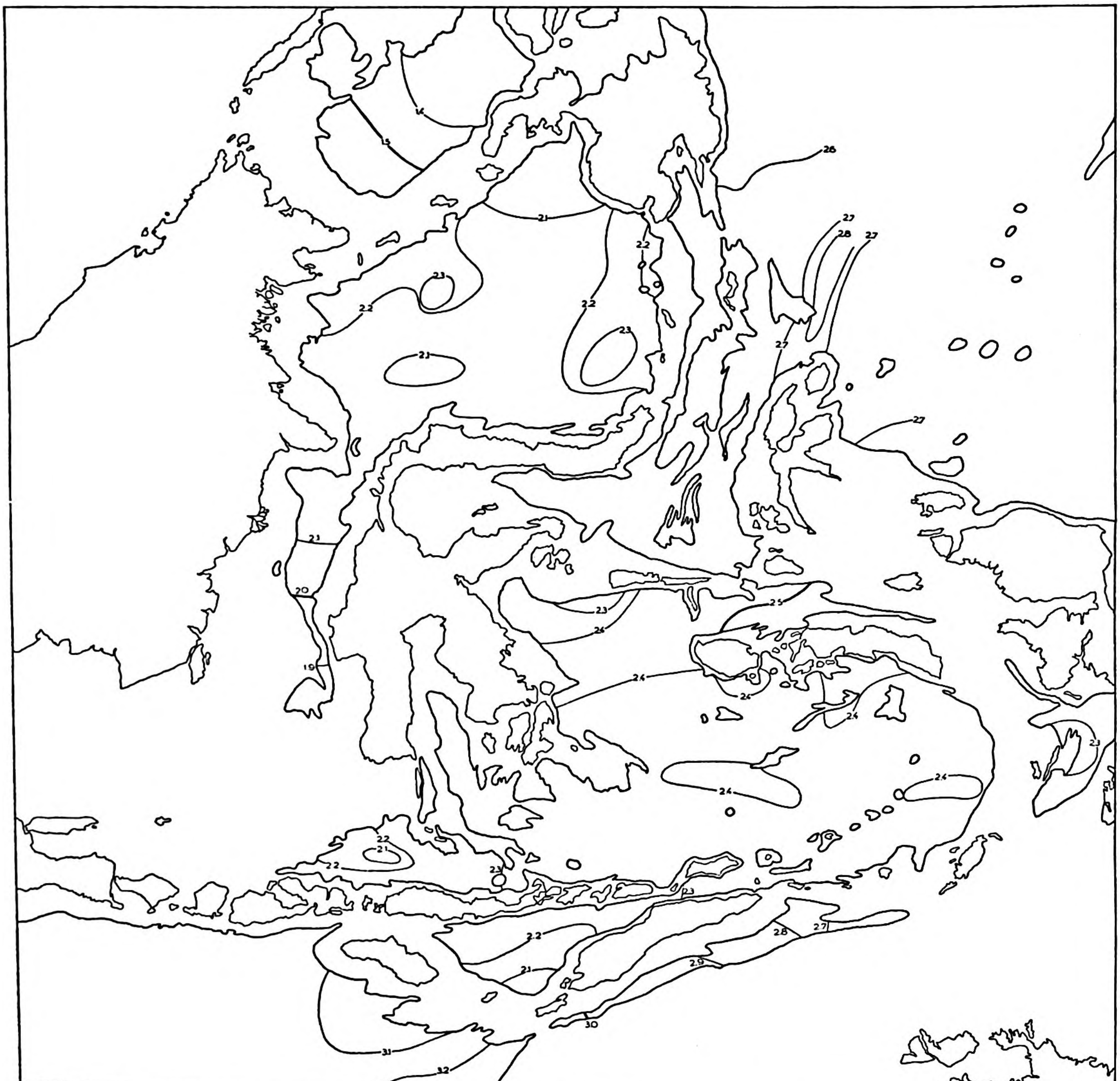

Fig. 23 I. Horizontal distribution of oxygen; 2000 m.

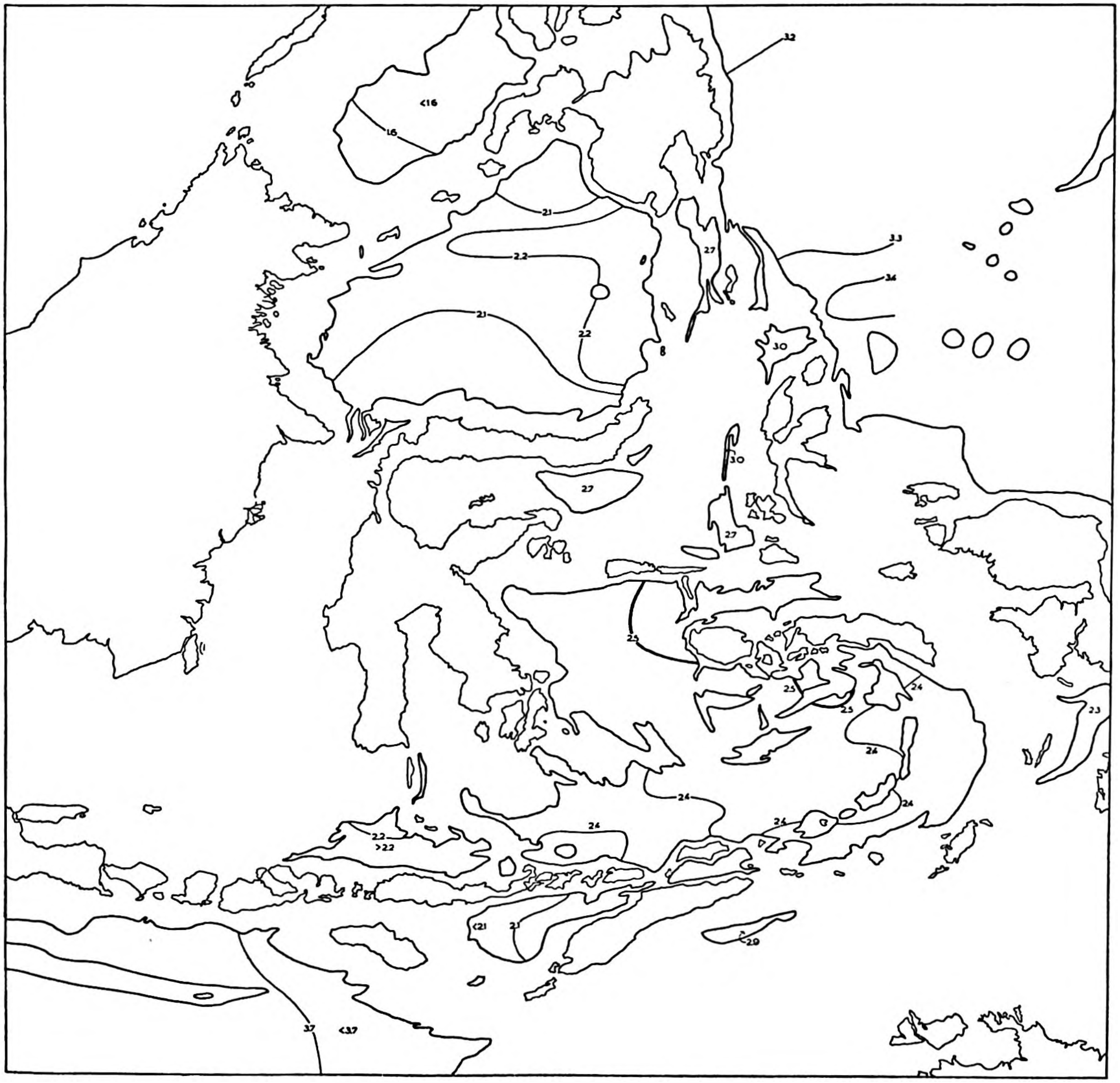

Fig. 23 K. Horizontal distribution of oxygen; 3000 m.

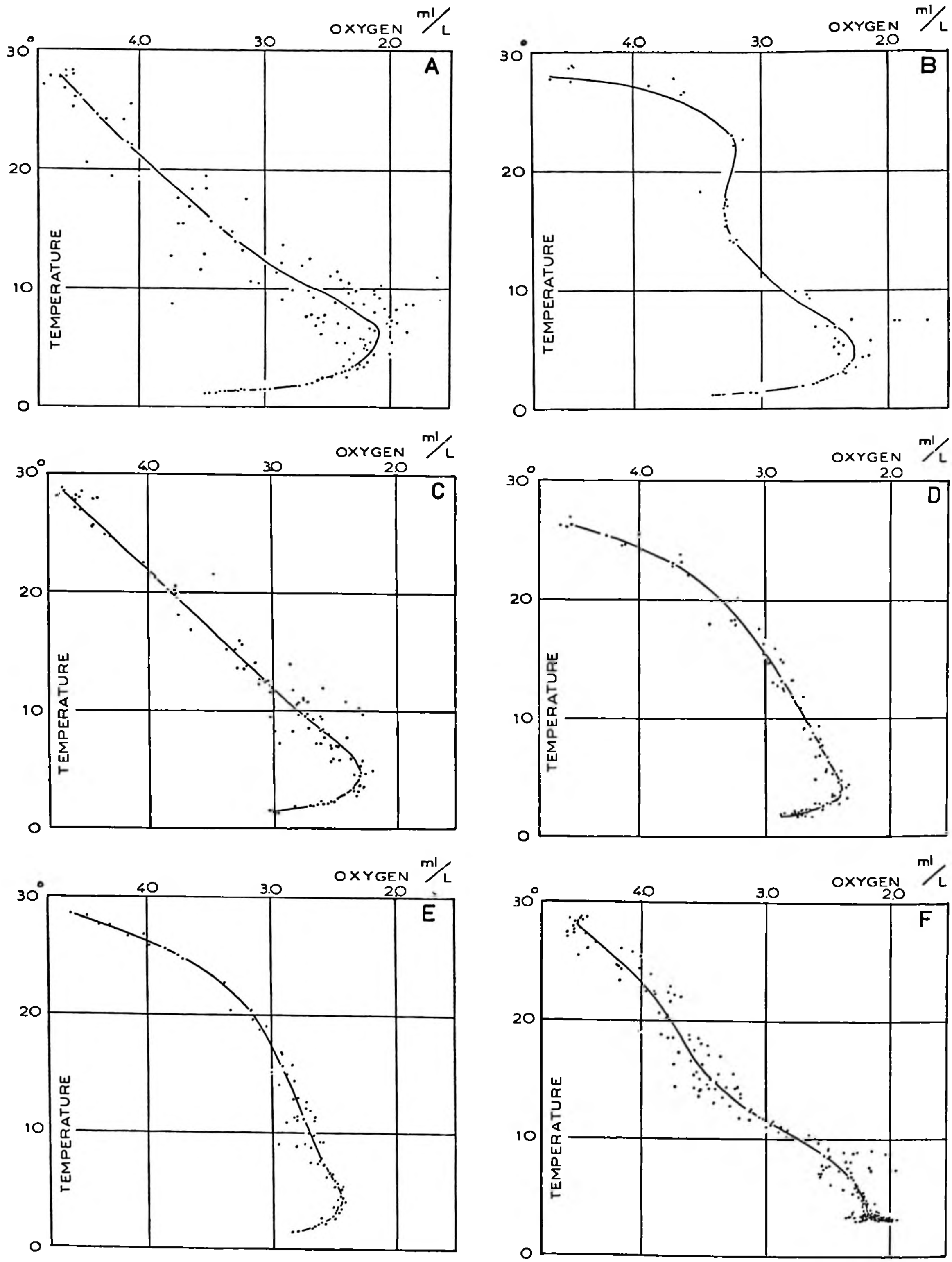

Fig. 24. Temperature-oxygen relations; compare table 9, p. 40

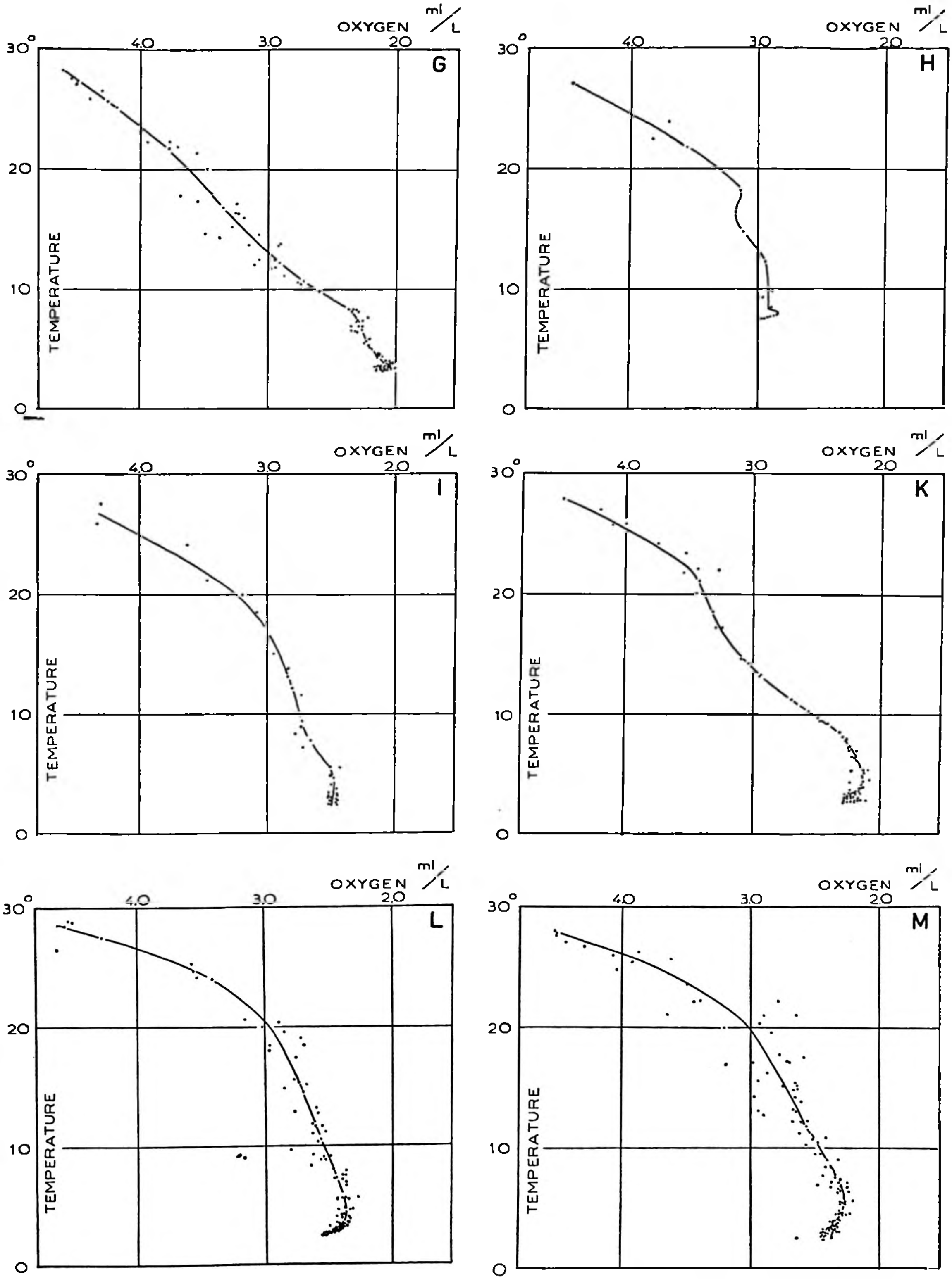

Fig. 24 (continued)

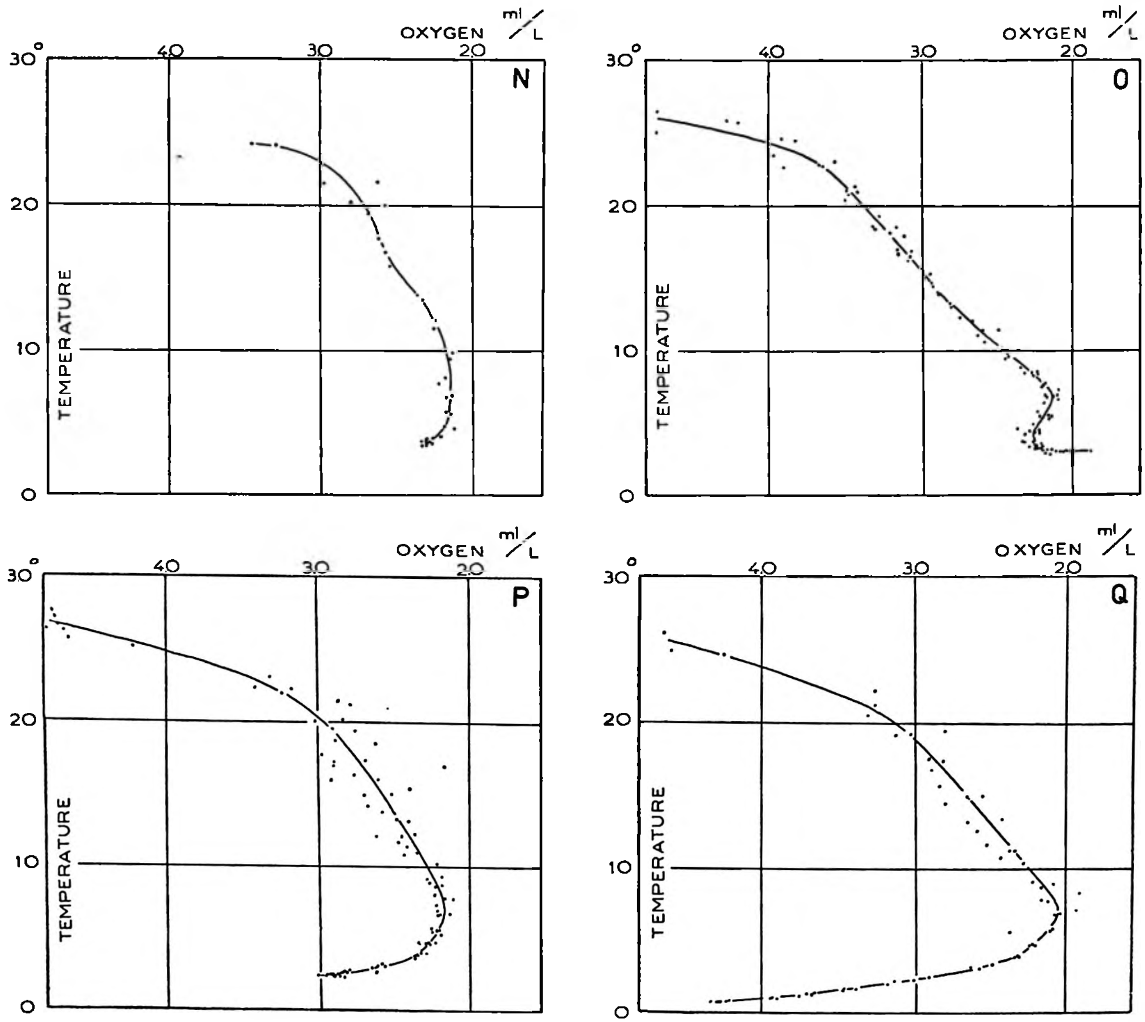

Fig. 24 (continued)